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Responses of an arctic landscape to Lateglacial and early Holocene climatic changes: the importance of moisture

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Abstract

Many of the physical and biological processes that characterize arctic ecosystems are unique to high latitudes, and their sensitivities to climate change are poorly understood. Stratigraphic records of land–surface processes and vegetation change in the Arctic Foothills of northern Alaska reveal how tundra landscapes responded to climatic changes between 13,000 and 8000 ^{14}C yr BP. Peat deposition began and shrub vegetation became widespread ca. 12,500 ^{14}C yr BP, probably in response to the advent of warmer and wetter climate. Increased slope erosion caused rapid alluviation in valleys, and *Populus* trees spread northward along braided floodplains before 11,000 ^{14}C yr BP. Lake levels fell and streams incised their floodplains during the Younger Dryas (YD) (11,000–10,000 ^{14}C yr BP). A hiatus in records of *Populus* suggest that its geographic range contracted, and pollen records of other species suggest a cooler and drier climate during this interval. Basal peats dating to the YD are rare, suggesting that rates of paludification slowed. Immediately after 10,000 ^{14}C yr BP, lake levels rose, streams aggraded rapidly again, intense solifluction occurred, and *Populus* re-invaded the area. Moist acidic tundra vegetation was widespread by 8500 ^{14}C yr BP along with wet, organic-rich soils. Most of these landscape-scale effects of climatic change involved changes in moisture. Although low temperature is the most conspicuous feature of arctic climate, shifts in effective moisture may be the proximate cause for many of the impacts that climate change has in arctic regions.

1. Introduction

The Arctic is important to global climate because its freshwater discharge affects North Atlantic circulation (Bacon, 1998), its snow cover influences global heat budgets (Foley et al., 1994), and its soils and wetlands are potential sources for greenhouse gases (Oechel and Vourlitis, 1996). Arctic climate is naturally variable (Overpeck et al., 1997) and is sensitive to human-induced changes (Maxwell, 1996). Although the Arctic is widely recognized as being a critical region for global climate change, the specific mechanisms by which climate change affects arctic landscapes are poorly understood.

Physical and biological processes in polar regions are assumed to be sensitive to rising temperature because the climate is cold, but in fact the critical processes remain obscure. Process studies suggest that moisture conditions are important in mediating between climatic changes and ecosystem responses in the Arctic (Hinzman and Kane, 1992). The dramatic changes occurring in water's physical properties around the freezing point have large effects on surface energy budgets (Kane, 1996). By thawing permafrost, rising temperatures can trigger changes in groundwater flow, surface drainage, and soil thermal regime (Nelson et al., 1993; Hinzman et al., 1996). Unfortunately, moisture conditions are poorly constrained in predictive models of arctic climate (Lynch et al., 1995; Moore et al., 1998; Bartlein et al., 1998; Kattsov and Walsh, 2000).

In this paper, we use the stratigraphic archives contained in fluvial, lacustrine, hillslope, and peat deposits to infer how surficial geology, soils, permafrost, and vegetation responded to rapid climatic changes occurring between 13,000 and 8000 ^{14}C yr BP during the Pleistocene to Holocene (P–H) transition on the North Slope of Alaska. Results indicate that changes in moisture were the proximate causes for many of these responses, which included feedbacks among permafrost, land–surface processes and vegetation, some of which are unique to arctic regions.

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2. Study area

Alaska's North Slope has two major physiographic divisions, the Arctic Foothills flanking the north side of the Brooks Range and the Arctic Coastal Plain lying further north between the Arctic Foothills and the Chukchi and Beaufort Seas (Fig. 1). Our 100×200 km study area is in the Arctic Foothills and crosses from the Brooks Range to the southern edge of the Arctic Coastal Plain. Both physiographic divisions of the North Slope are described here because their environmental histories are closely interrelated.

The Arctic Foothills are east–west trending ridges of carbonate bedrock that protrude from tundra-covered plains (Grantz et al., 1994). Permafrost is continuous north of the Brooks Range and reaches hundreds of meters in thickness (Ferrians, 1994). Much of the Arctic Foothills region has never been glaciated. Areas near the Brooks Range, including the Mesa archaeological site (Kunz and Reanier, 1994), were last glaciated during the Tertiary and early Pleistocene. During the Last Glacial Maximum (LGM), glaciers in the Brooks Range terminated along the northern range front (Hamilton, 1986).

The Arctic Coastal Plain is underlain by a broad, low relief bedrock surface that dips gently seaward. Over the last 3 million years, the sea has repeatedly transgressed and regressed across this surface, leaving a veneer of unconsolidated and interfingering marine and non-marine deposits (Dinter et al., 1990). Prominent among the non-marine deposits are sands and silts derived from river channels and deltas. During dry intervals in the Pleistocene, these sediments were re-worked by the wind into extensive dune fields and loess belts (Carter et al., 1987; Carter, 1988). Immediately north of the study area, the Arctic Coastal Plain is underlain by the now stable Ikpikpuk Dunes, which formed a 12,000-km^2 sand sea during the LGM and was partly reactivated several times during the Holocene (Carter, 1981, 1993; Dinter et al., 1990; Galloway and Carter, 1993).

Marked north–south gradients in climate occur across the North Slope. July mean temperature increases from 4° C at Barrow to 12° C at Toolik Lake near the Brooks Range front (Zhang et al., 1996). Mean annual precipitation increases inland from 200 mm at Barrow to 320 mm at Toolik Lake. Throughout the region, about half of the precipitation falls as snow, which persists on the ground for more than 8 months of the year. Rainfall increases over the course of the summer with maxima accompanying cyclonic storms in July, August, and September (Kane et al., 1992). Many of these storms cross the Brooks Range from the Bering Sea (Moritz, 1979).

During summer, most of the North Slope exists in a state of waterlogged aridity. Potential evapotranspira-

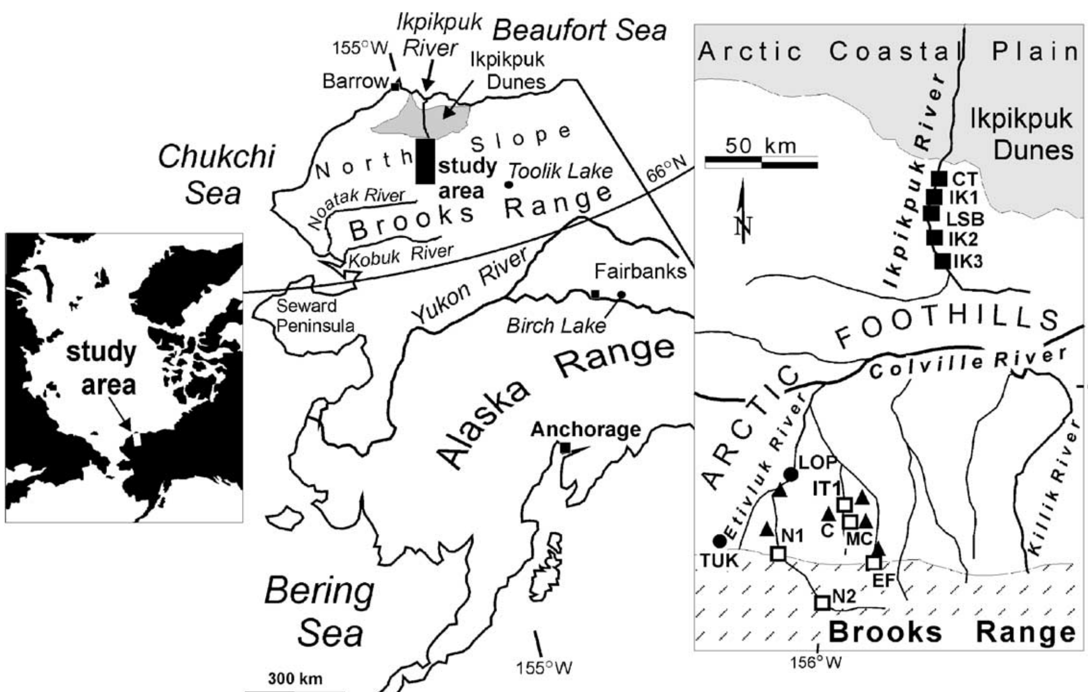

Fig. 1. The Arctic Foothills lie between the Brooks Range and the Arctic Coastal Plain on Alaska's North Slope. In the right panel, squares show locations of fluvial sections. N1 is Nigu 1, N2 is Nigu 2, EF is on the East Fork of the Etivluk River, MC is Mesa Creek section, and IT1 is on Iteriak Creek just downstream of the Mesa archaeological site. Along the Ikpikpuk River, CT is Cottonwood Bend, IK1 is Ikpikpuk 1, LSB is Little Supreme Bluff, IK2 is Ikpikpuk 2, and IK3 is Ikpikpuk 3. LOP is "Lake of the Pleistocene." Triangles represent solifluction sections; C is Cobra Gulch. TUK is Tukuto Lake. The portion of the Arctic Coastal Plain shown here includes part of the stabilized Ikpikpuk Dunes.

tion equals or exceeds total annual precipitation (Rovansek et al., 1996), actual evapotranspiration exceeds annual runoff (Hinzman et al., 1996), and the region is classified as semi-arid by the Thornthwaite method (Patric and Black, 1968; Newman and Branton, 1972). Nonetheless, soils at many sites remain saturated throughout the summer because water tables are perched on the frozen, ice-rich substrate, and water is concentrated at the ground surface. This situation can persist because evapotranspiration is greatest in early summer before active layers fully thaw, and precipitation in late summer re-charges soil moisture at a time when evapotranspiration is low (Hinzman et al., 1996; Zhang et al., 1997).

Tundra covers the North Slope; however, a major vegetation boundary lies along the northern edge of the Arctic Foothills (Walker et al., 1998). In the Arctic Foothills, most of the vegetation is moist acidic tundra (*Sphagno-Eriophoretum*) dominated by dwarf shrubs (*Betula nana*, *Ledum palustre*, *Salix planifolia pulchra*), tussock sedges (*Eriophorum vaginatum*) and acidophilous mosses, among which *Sphagnum* species are prominent. An important point for the interpretation of pollen records is that ericaceous shrubs, *Sphagnum* moss, and *Rubus chamaemorus* (cloudberry) are characteristic of moist acidic tundra vegetation today (D. Walker et al., 1989, 1998, 2001; M. Walker et al., 1994; Shaver et al., 1996; Walker and Walker, 1996).

Moist acidic tundra vegetation typically is underlain by peaty organic horizons, leaving little mineral soil exposed at the ground surface (Bockheim et al., 1998). In the Arctic Foothills today, much of the landscape is blanketed under 5–40 cm of peat (Everett and Brown, 1982; Ping et al., 1998). Peat, soil material containing $>30\%$ organic matter (Heathwaite et al., 1993), insulates the ground and creates active layers that are only 30–40 cm thick (Bockheim et al., 1998). Organic surface horizons acidify the soil and isolate plant roots from the mineral soil. Plant species diversity is relatively low in moist acidic tundra, plant tissues generally have low nutrient levels, and the dominant plant species are strongly defended by anti-herbivore, secondary compounds (Walker et al., 2001). These characteristics, coupled with the difficulty of walking through swampy tussocks, make moist acidic tundra a relatively poor habitat for large herbivores.

On the Arctic Coastal Plain, the dominant vegetation type is moist non-acidic tundra (*Dryado-integrifoliae–Caricetum bigelowii*), which is dominated by non-tussock sedges (*Carex bigelowii*, *C. membranacea*, and *Eriophorum triste*), prostrate shrubs (*Dryas integrifolia*, *Salix arctica*, *S. reticulata*, and *Arctuous rubra*), and minerotrophic moss taxa (Walker et al., 1998). In comparison to the acidic tundra prevalent in the Arctic Foothills, moist non-acidic tundra has higher soil pH (~ 7), thinner and more discontinuous organic surface horizons, deeper active layers, and a higher degree of frost disturbance (Bockheim et al., 1998; Walker et al., 2001). Plant species diversity is higher, and many plants are disturbance-adapted species with higher tissue-nutrient levels and fewer anti-herbivore compounds. Modern caribou calving grounds are located mainly in areas of non-acidic tundra (Walker et al., 2001).

Soil development on the North Slope leads towards paludification, the accumulation of waterlogged organic material on previously well-drained terrain (Bockheim et al., 1998). The waterlogging that accompanies peat accumulation influences vegetation distribution throughout the region (Walker and Walker, 1996; Walker et al., 1998), and *Sphagnum* mosses play a key role in paludification on the North Slope by acidifying soils, lowering soil-nutrient levels, and retaining water. In poorly drained areas, active layers are typically only 25 cm thick as a result of insulation by organic surface layers (Everett and Brown, 1982; Kane, 1996). In contrast, summer thaw on well-drained slopes with thin organic surface horizons can penetrate to depths of 1 m (Kane, 1996). Thick organic surface layers restrict frost heaving and so limit the transfer of nutrient-rich, mineral material to the soil surface (Walker et al., 2001). Organic soil horizons affect floodplain dynamics by retarding slope erosion and limiting the input of sediments to streams (Hinzman et al., 1996).

Floodplains play key roles in North Slope ecosystems. By contrast with much of the surrounding terrain, floodplain soils have deep active layers, lack thick organic horizons, and are often well drained (Ping et al., 1998). They are sites of relatively high primary productivity and plant-species diversity (Shaver et al., 1996; Walker et al., 2001). The unvegetated portions of floodplains are sources of loess, whose deposition has pervasive effects on soils and vegetation located downwind of, and adjacent to, floodplain margins (Walker and Everett, 1991). Loess and sand that are deposited on soil surfaces retard soil acidification, which reduces the rate of organic-matter accumulation and thereby maintains relatively deep and well-drained active layers.

The major runoff event of the year on the North Slope is the snowmelt flood (Arnborg et al., 1966; Carter et al., 1987; Kane, 1996); however, frozen soils restrict erosion during snowmelt, and summer rainstorms probably cause most erosion on slopes. This is also true for the beds of smaller streams where bottom-fast ice and frozen sediments armor channels during snowmelt (Scott, 1978). Where channels impinge onto higher terrain, a large part of lateral channel migration is due to thermal erosion of ice-rich permafrost (Carter et al., 1987). Most streams have wandering, gravel- and cobble-bedded channels today, though braided reaches exist in areas of *aufeis* accumulation, and low-order streams are often beaded or straight channeled between banks of peat.

Trees are absent from the North Slope except for several, widely separated stands of *Populus balsamifera* (balsam poplar) growing on floodplains in the Arctic Foothills (Edwards and Dunwiddie, 1985). With a mean July temperature of 10–12° C, the Arctic Foothills lie near latitudinal treeline today (Hopkins, 1959; MacDonald et al., 2000).

3. Previous studies

3.1. Glacier fluctuations

Glaciers in the Brooks Range retreated from LGM before 14,000 ^{14}C yr BP (Hamilton and Porter, 1975; Hamilton, 1986) and underwent minor re-advances near valley heads between 13,000 and 11,500 ^{14}C yr BP, perhaps in response to increasing winter snowfall (Hamilton, 1986). Neoglaciation began in the Brooks Range perhaps as early as 7600 yr ago, though widespread advances did not occur until after 5000 yr ago (Calkin, 1988). Holocene glaciation in the Brooks Range was confined to cirques and valley heads.

3.2. Paleohydrology

Streams and their floodplains react sensitively to climatic changes (Porter et al., 1992; Macklin, 1999; Knox, 2000), though differences in climate, vegetation, soil conditions, and watershed position can cause large variations in these responses (Knox, 1983). We know that climatic changes during the P–H transition triggered sweeping changes in the river systems of northwestern Europe by altering effective moisture, vegetation cover, sediment supply, and permafrost conditions (Frenzel, 1995; Collins et al., 1996; Huisink, 1997; Tebbens et al., 1999; Bridgland, 2000); however, there are no previous studies of how climate change affected fluvial geomorphology on Alaska's North Slope during the P–H transition.

Hamilton (1980, 1982, 1984) mapped fluvial terraces along the northern front of the Brooks Range, related them to glacial advances of Tertiary and Pleistocene age, and described several stratigraphic sections, ^{14}C dates from which are incorporated in Appendix C. Dinter et al. (1990) mapped fluvial deposits of Cenozoic age on the Arctic Coastal Plain and related episodes of terrace building to times of sea-level high stands. Holocene and Pleistocene alluvial fills and terraces along streams on the North Slope have not been studied in detail, though scattered ^{14}C dates indicate that many formed between 14,000 and 8000 ^{14}C yr BP (Nelson and Carter, 1987; Dinter et al., 1990). The Pleistocene glaciation of the Brooks Range created a complex succession of glacier-dammed lakes with associated fluvial terraces and shorelines in the upper Noatak valley (Hamilton, 2001). Glaciation was accompanied by alluvial and aeolian activity in the Kobuk valley south of the range (Ashley and Hamilton, 1993; Hamilton and Ashley, 1993). Hamilton et al. (1988) concluded that extensive, but presently inactive, gulley networks in loess slopes near Fairbanks were cut during the last several millennia of the Pleistocene in response to a sudden increase in runoff.

In a well-dated, lake-level record from the P–H transition at Birch Lake in interior Alaska (Fig. 1), water levels are inferred to have risen rapidly between 12,700 and 12,200 ^{14}C yr BP, thereafter falling back to low levels (Abbott et al., 2000). A second transgression occurred sometime between 10,600 and 10,000 ^{14}C yr BP. Between 8800 and 8000 ^{14}C yr BP, the lake rose to reach its modern (overflow) level.

Epstein (1995) measured deuterium in ^{14}C-dated *Salix*-wood samples from the now inactive Ikpikpuk Dunes on the Arctic Coastal Plain. His results suggest that major fluctuations in water availability, and/or in evaporative source areas for precipitation, occurred between 12,000 and 8000 ^{14}C yr BP. Cellulose LD values were highly negative during the Younger Dryas Chronozone ($\sim$11,000–10,000 ^{14}C yr BP), suggesting more droughty conditions and/or a more distant evaporative source. As discussed in detail below, changes in soil moisture triggered episodic reactivations of the Ikpikpuk Dunes during the Lateglacial and Holocene (Carter, 1993; Galloway and Carter, 1993).

3.3. Paleobotany

The vegetation history of Alaska's North Slope during the P–H transition is known only in outline due to the reconnaissance nature and poor dating control of pioneering studies and because of methodological problems inherent to tundra-zone palynology. Some previous studies are based on discontinuously deposited, silt and peat sediments of Holocene age (Livingstone, 1957; Colinvaux, 1964a, b; Walker et al., 1981; Eisner and Peterson, 1998). Where Pleistocene- and Holocene-aged lake sediments have been analyzed, the sampling intervals are too wide to yield temporal resolutions better than several millennia (Eisner and Colinvaux, 1990, 1992). More detailed pollen records from lake sediments are available from far northwestern Canada (Cwynar, 1982; Ritchie et al., 1983), though dating resolution is limited there also. Pollen records from lakes on the southern slopes of the Brooks Range (Brubaker et al., 1983; Anderson, 1985, 1988) span the P–H transition, but more recent studies (Oswald et al., 1999) suggest their dating control is problematic

Accurate dating control is difficult to achieve in arctic lakes because low temperature and waterlogged soils allow abundant organic material to escape decay and to become incorporated in younger sediments (Schell, 1983; Abbott and Stafford, 1996). Bulk samples of lake sediment can be contaminated by old carbon derived from soils, peat, and/or carbonate bedrock within the watershed (Oswald et al., 1999). ^{14}C-dating of paired samples reveals that bulk-sediment samples can be centuries to millennia older than the associated macrofossils of terrestrial plants (Abbott and Stafford, 1996; Abbott et al., 2000).

Tundra pollen records can be difficult to interpret for several reasons. Some abundant and presently diverse groups of tundra plants are impossible to identify to species using pollen characteristics. Many tundra species are ecological generalists with wide geographical ranges, limiting their usefulness in paleo-environmental interpretations (Colinvaux, 1964a; Anderson et al., 1994). Pollen accumulation rates (PARs) are of limited use for interpreting vegetation changes during the P–H transition because sedimentation rates are typically slow, variable, and, as just described, difficult to date accurately. In sediments older than 10,000 ^{14}C yr BP, PAR estimates are complicated further by uncertainties in the calibration of ^{14}C yr to calendar years.

With these cautions about taxonomic/ecological resolution and age control in mind, the vegetation history of the Arctic Foothills from $\sim$19,000 ^{14}C yr to present can be described as follows. The work of Oswald et al. (1999) at Tukuto Lake (Fig. 1) is emphasized because it is the most carefully executed and best-dated lake-core record near our study area.

Vegetation history in the area around Tukuto Lake followed the same sequence of herb-, birch-, and alder-pollen zones that is documented in other parts of eastern Beringia (Anderson and Brubaker, 1994; Edwards and Barker, 1994). In levels dating to the LGM, Poaceae (grass) is the most abundant pollen taxon (30–50%), with less abundant Cyperaceae (sedge, $<10\%$), and a variety of minor herb taxa. Oswald et al. (1999) suggest that vegetation in the Arctic Foothills at this time was xeric tundra. At levels dated sometime between the LGM and $\sim$13,000 ^{14}C yr BP, Poaceae percentages decline and Cyperaceae and Salix (willow) become the dominant pollen taxa. Vegetation in this transition period probably was similar to the moist non-acidic tundra now widespread on the Arctic Coastal Plain. The transition from xeric, steppe–tundra to mesic, shrub tundra was completed sometime after 12,830 ^{14}C yr BP. At the level of this date, *Betula* (birch) pollen, probably that of *Betula nana/glandulosa* (dwarf and shrub birch), reaches values $>20\%$ in the pollen diagram (Oswald et al., 1999). Shrub tundra vegetation is indicative of increased summer warmth and increased effective moisture (Cwynar, 1982; Chapin et al., 1995).

The position of P–H transition is poorly constrained in the Tukuto Lake core between levels dated to 12,800 and 7300 ^{14}C yr BP. Only 15 samples from this 5500 ^{14}C-yr interval were analyzed for pollen, so vegetation history during the transition remains obscure. It is unclear from the plot of total PAR where in the lithostratigraphy the P–H transition actually occurs. Pollen of Ericales (heaths) and *Rubus chamaemorus* first appear at levels dated between 12,800 and 7300 ^{14}C yr BP, and Oswald et al. (1999) suggest that moist acidic tundra became widespread during this period.

A diffuse peak in *Populus* (cottonwood) pollen occurs in the Tukuto Lake core between limiting dates of 12,800 and 7300 ^{14}C yr BP. Elsewhere in northern Alaska, a *Populus*-pollen subzone is dated between 11,000 ^{14}C yr BP and 8000 ^{14}C yr BP; though the timing of its beginning and end vary by as much as 3000 ^{14}C yr in different lakes, and there is no apparent geographical pattern to this variability (Anderson, 1988; Bartlein et al., 1995; Szeicz and MacDonald, 2001). Questions about the accuracy of the bulk-sediment ^{14}C dates make it unclear whether or not the *Populus* subzone was synchronous across the region. Radiocarbon dates on *Populus* wood at sites now beyond the present range of this species suggest that *Populus* trees had expanded into areas beyond their present range limits prior to 11,000 ^{14}C yr BP (references in Appendix A). Prior to this study, 15 dates on extralimital *Populus* wood from northern Alaska and northwestern Canada had been reported. Three of these dates fall between 11,500 and 11,100 ^{14}C yr BP, while the remainder fall between 9900 and 7200 ^{14}C yr BP.

Alnus (alder) pollen appears in the Tukuto Lake record in levels younger than 7300 ^{14}C yr BP. The arrival of alder probably indicates further increases in effective moisture (Oswald et al., 1999). The *Alnus* pollen zone represents a period of general vegetational stability near Tukuto Lake that continues to the present. Today the vegetation near the lake is dominated by moist acidic tundra with *Betula*, *Alnus*, *Salix*, Ericales, and Cyperaceae dominating the pollen rain, and *Sphagnum* the most abundant spore type (Oswald et al., 1999).

3.4. Fossil beetles

Beetle remains from northern Alaska and the continental shelf of the Chukchi Sea suggest that summer and winter temperatures were depressed by 2° and 8° C, respectively, during the LGM (Elias et al., 1996; Elias, 2000, 2001). Beetle remains from western Alaska indicate that temperatures had risen from ice-age to modern levels by 12,500 ^{14}C yr BP. A peak in warmth occurred $\sim$11,000 ^{14}C yr BP and was followed by a possible cooling between 11,000 and 10,000 ^{14}C yr BP (Elias, 2000). A beetle fauna associated with *Populus* wood dating to 9430 ^{14}C yr BP at a site on the Ikpikpuk River indicates well-drained soils and a July mean temperature 2–3°C higher than today (Nelson and Carter, 1987).

4. Methods

Stratigraphic sections were prepared by clearing slumped material to expose undisturbed, frozen sediments. Basal peats were recovered from permafrost using a power auger. At each drill site, coring was continued until breaking a cutter tooth on rock. Particle size (Udden-Wentworth scale) of mineral sediments in Lake of the Pleistocene was determined by the pipette method; coarser sediments were analyzed by dry sieving. In both procedures, organic material and carbonates were removed using H_2O_2. Organic-matter content was estimated after loss on ignition at 500°C. Total inorganic carbon content was measured by subsequent heating of the same samples to 1000°C (Dean, 1974). For pollen analysis, subsamples of 1 or 2 cm^3 of sediment were prepared for analysis following standard procedures for organic-poor sediments (PALE, 1994). Pollen residues were mounted in silicon oil and counted at 400x and 1000x magnification. At least 300 grains of terrestrial pollen were counted at each level. The pollen percentages for tress, shrubs, and herbs were based on this sum. Percentages of terrestrial spores were based on the terrestrial pollen plus terrestrial spore sum, and percentages of aquatic spores were based on the terrestrial pollen plus the aquatic and spore sum. The mean number of indeterminate grains per counted level was 4.30 (s.e. = 0.35). ^{14}C dates are calibrated using Calib4 (Stuiver et al., 1998).

Fluvial sections were described by mapping sedimentary units and bounding surfaces onto photographic mosaics of cutbank exposures. Stream terraces were mapped using 1:60,000- and 1:24,000-scale aerial stereo photography, and their altitudes estimated using topographic maps with contour intervals of 10–20 m and a surveying altimeter with an accuracy of ± 2 m. Paleoflow directions were inferred from the orientations of bedform slipfaces, the strikes of scour channels, and occasionally from the orientation of woody debris. Plant macrofossils were extracted for AMS-radiocarbon dating by washing sediments through 500- and 150-micron sieves, examining residues under a dissecting microscope, and identifying plant parts using modern reference material. Wood identifications were made by the US Forest Service Wood Products Laboratory in Madison, Wisconsin. *Populus balsamifera* leaves were identified at the University of Alaska Herbarium by C. Parker and D. Murray.

5. Results and discussion

5.1. Lake of the Pleistocene

Lake of the Pleistocene (LOP) is a drained lake basin whose sediments contain a lengthy record of lake-level and vegetation changes. The basin formed as an alas valley (Yershov, 1998) in glacial-outwash terraces of middle Pleistocene age (Hamilton, 1984) (Fig. 2). Intermittent overflow probably occurred northward over a series of low sills formed by the alluvial fans of streams such as Rudivik Creek. Sediments accumulated in LOP from >40,000 until ~5000 ^{14}C yr BP (Appendix B), when the basin was breached by lateral erosion of the Etivluk River. The deepest part of the basin still contains Nikivlik Lake, a 3-m deep remnant of the formerly larger lake. Today, several hundred meters of the ancient lakebed (Fig. 3) are cross-sectioned by the Etivluk River, revealing a detailed lithostratigraphic sequence containing abundant twigs suitable for AMS-^{14}C dating. The LOP sections lack the sedimentary facies diagnostic of thaw-lake basins (Hopkins and Kidd, 1988; Murton, 1996), indicating that the sediments studied here have not been disturbed by thermokarst processes.

5.1.1. Water-level history in LOP

Striking features of the Holocene portions of the LOP sections are 0.5–2 cm thick layers of flat-lying willow leaves and twigs, which are laterally continuous for 20–100 m and occur sporadically in vertical section (Fig. 3). These plant-debris layers contain sand and occasional granules and sometimes comprise the basal portions of

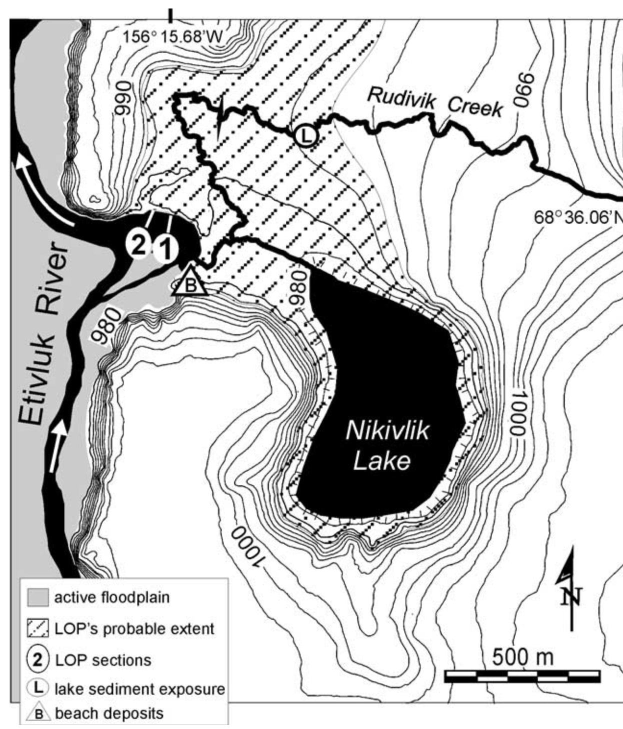

Fig. 2. Part of the former drainage basin of LOP. The 2 m contour interval comes from a laser-transit survey. The extent of the former lake is based on an exposure of lake sediments along Rudivik Creek and a sandy beach deposit on the southern shore. Solifluction has obscured the ancient shorelines throughout the basin.

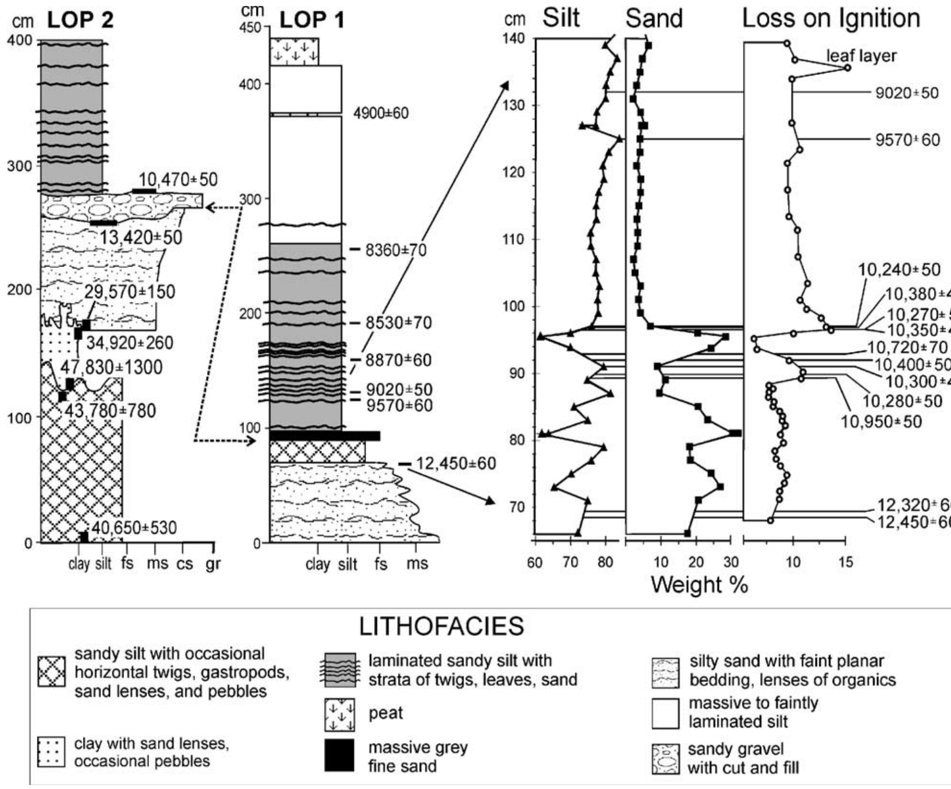

Fig. 3. Stratigraphy of LOP. Section LOP2 is 50 m west of LOP1 and closer to the basin's margin. "Fs" is fine sand, "ms" is medium sand, "cs" is coarse sand, and "gr" is gravel. All ^{14}C dates were obtained by the AMS method on willow and birch twigs.

normally graded beds capped by silts and clays. We interpret these beds as recording ancient sedimentation limits, the border zones between wave-scoured shallows and deeper, still water where fine sediments are deposited (Dearing, 1997; Verschuren, 1999). Similar plant-debris layers occur today in Nikivlik Lake, where we studied them using a surface-sediment sampler and by wading barefoot.

Presently in both summer and winter, abundant plant debris is blown from lee-slope willow thickets into Nikivlik Lake. After initial deposition near shore, waterlogged twigs and leaves are transported into deeper water during high-wave events, where they accumulate at depths just below wave base (cf., Sly, 1978). As the water calms, suspended silt and clay settle out on top of the plant debris. Today, zones of multi-layered, muddy twigs and leaves occur on the bottom of Nikivlik Lake in water depths between 0.2 and 1 m. If water depth increased and the lake expanded, the zone of abundant plant debris would migrate landward. In vertical sections through the silt- and clay-rich Holocene sediments of LOP, units with few or no plant-debris layers indicate relatively deep water, while those with numerous plant-debris layers indicate shallow water.

Changes in particle size, organic content, and debris-layer frequency (Fig. 3) indicate that water levels fluctuated markedly in LOP. Low sedimentation rates and sandy sediments suggest that the lake was intermittently dry during the LGM. By 12,450 ^{14}C yr BP, water level had risen sufficiently that sandy silts containing gastropods and algal cells were deposited. In the LOP1 section, organic content increased shortly before 11,000 ^{14}C yr BP then declined to low values between 10,900 and 10,200 ^{14}C yr BP during the Younger Dryas (YD). ^{14}C dates within this interval show the age reversals that are characteristic of the YD Chronozone (Björck et al., 1996). In outcrop, the YD corresponds to 3–5 cm of fine sand that is laterally continuous for >100 m, and which probably records a low lake stand. Alternately, the sand layer could record a turbidity flow resulting from a landslide in the drainage basin. The latter hypothesis is rejected because the sequence of ^{14}C dates during the YD Chronozone in LOP1 follow a characteristic pattern found at other sites in the northern hemisphere (Hajdas et al., 1998), indicating that sedimentation proceeded slowly but continuously over the ~ 1000 ^{14}C-yr duration of the YD.

In the LOP2 section, which lies 50 m closer to the basin margin, 10-20 cm of medium sand with pebble lags and an erosive lower contact probably represents a beach formed during the YD low stand of the lake (Fig. 3). Altitudinal differences between the two sections are obscured by deformation caused by epigenetic ice wedges, which formed in the sediments after lake drainage ~ 5000 ^{14}C yr BP. Deposition of organic silts and clays resumed after $\sim 10{,}200$ ^{14}C yr BP as water levels rose. The distribution of plant-debris layers suggests that the lake deepened between 10,000 and 9600 ^{14}C yr BP, but then shallowed until ~8600 ^{14}C yr BP, after which it deepened again.

5.1.2. Vegetation history during the Pleistocene–Holocene (P–H) transition in LOP

The lowest two samples in the LOP1 section are dominated by Cyperaceae pollen (40–60%) with minor amounts of Poaceae (grass) (8%) and *Juniperus* (juniper) (1%) pollen (Fig. 4). Intensive sampling of the section begins at the 68-cm level between two twigs that yield concordant ages of 12,450 and 12,320 ^{14}C yr BP. At this level, *Betula* (birch) pollen has already increased to 20%, which is its percentage in modern pollen-rain samples from the northern range limit of *B. nana/glandulosa* in Alaska (Anderson and Brubaker, 1986, 1994). We infer that *Betula* shrubs had become widely established around LOP at or shortly before 12,400 ^{14}C yr BP. *Populus* pollen reach 4%, this taxa's all-time high in the diagram, at the level of the 12,320 yr date, and cells of the alga *Pediastrum* begin an abrupt rise at the same level, suggesting the LOP basin had filled with water. All these changes are suggestive of an increase in effective moisture within the several centuries preceding 12,400 ^{14}C yr BP.

Several subtle changes occur in the LOP pollen diagram between the 87- and 95-cm levels in the part of the section corresponding to the YD. *Populus* pollen disappear immediately above the level dated to 10,900 ^{14}C yr BP and do not re-appear in the diagram until slightly above the 100-cm level, which lies between limiting ages of 10,240 and 9570 ^{14}C yr BP (Fig. 4). *Populus* pollen decays readily, and its absence from the sand unit could be an artifact of preservation; however, *Populus* is also absent from lowest 5 cm of silty sediments immediately overlying the sand unit. We hypothesize that the temporary disappearance of *Populus* indicates a reduction in the extent of its preferred floodplain habitat due to changes in fluvial regimes (see below) and perhaps because of cooler, drier summers.

Changes in the percentages of several other taxa hint at drier conditions during the YD. A peak in *Juniperus* pollen percentages occurs at levels dating to 10,950 ^{14}C yr BP. Today, *Juniperus* grows at xeric sites throughout Alaska. The cause of a brief spike in *Betula* pollen at levels corresponding to the YD Chronozone remains unexplained. *Artemisia* (sage) and Poaceae pollen and *Selaginella sibirica* (spikemoss) spores reach their highest abundance in the entire record during the YD Chronozone. *Selaginella sibirica* is common today on windblown, south-facing slopes (Walker et al., 1989). *Pediastrum* briefly reached high abundance during the YD, possibly suggesting lower water levels (Anderson and Brubaker, 1986; Bigelow and Edwards, 2001). A brief decline in Cyperaceae percentages and increases in *Artemisia*, Poaceae, *Juniperus*, *Selaginella*, and *Pediastrum* are all consistent with drier conditions during this interval. Similar patterns of change by some of these same taxa have been interpreted as evidence for drier and cooler conditions during the YD in the Alaska Range (Bigelow and Edwards, 2001).

Changes in pollen and spore percentages at levels dating between 10,240 and 8530 ^{14}C yr BP are suggestive of increasing effective moisture. Ericales percentages rise gradually in levels younger than 12,320 ^{14}C yr BP, but they do not exceed 5%, their percentage in the modern pollen rain on the North Slope (Anderson and Brubaker, 1986), until the level dated to 9570 ^{14}C yr BP.

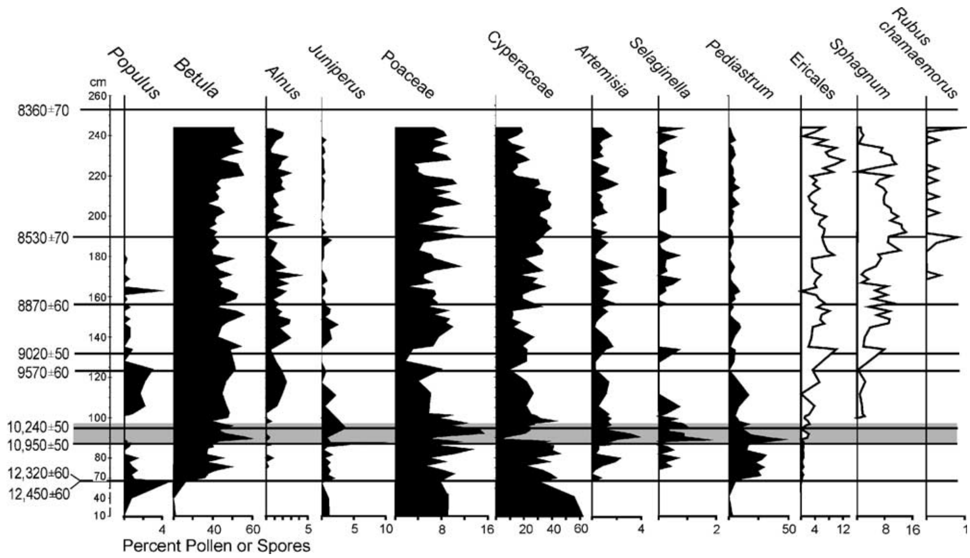

Fig. 4. Percentage pollen and spore diagram from LOP1. White curves show taxa associated with paludification. ^{14}C dates from LOP1 section are shown in Fig. 3. The YD Chronozone is shaded.

Sphagnum first reaches its modern percentages (10%) near the level dated to 9020 ^{14}C yr BP. Both Ericales and *Sphagnum* are indicator taxa of moist acidic tundra (Walker et al., 1998). Another indicator of this vegetation type is *Rubus chaemaemorus* (cloudberry), whose pollen first appears in the diagram at levels dating to between 8870 and 8530 ^{14}C yr BP.

Slight declines occur in *Juniperus,* Poaceae, *Artemisia*, *Selaginella*, and *Pediastrum* percentages in samples post-dating 10,240 ^{14}C yr BP (Fig. 4). *Populus* disappears entirely at levels younger than 8500 ^{14}C yr BP. *Alnus* pollen first reaches values of several percent between 10,240 and 9570 ^{14}C yr BP, a percentage maintained through the remainder of the diagram. Such low percentages make it difficult to ascertain when *Alnus* first arrived near LOP. We conclude that the modern tundra vegetation, accompanied by its poorly drained, organic soils, was probably established in the study area between 9000 and 8500 ^{14}C yr BP.

5.2. Extralimital Populus trees

New and previously obtained ^{14}C dates on *Populus* wood and leaves from northern Alaska and northwestern Canada suggest this species expanded into areas beyond its present range limits twice during the latest Pleistocene and earliest Holocene. We dated two *Populus* logs and two *Populus balsamifera* leaves from fluvial sediments along the Ikpikpuk River. The two logs and one of the leaves dated between 9270 and 9720 ^{14}C yr BP, and the other leaf dated to 10,940 ^{14}C yr BP (Appendix A). Four previously reported dates on cottonwood from the Ikpikpuk River range from 8710 to 9540 ^{14}C yr BP (Nelson and Carter, 1987). Of the ten samples of extralimital *Populus* wood reported by Hopkins et al. (1981), three dated to between 11,100 and 11,500 ^{14}C yr BP, and the remainder between 7270 and 9940 ^{14}C yr BP. One of the pre-11,000 year dates was from the Nigu River, probably from the Nigu 1 section described below. These results suggest that *Populus* expanded its range into the present tundra region twice, first between 11,500 and 10,900 ^{14}C yr BP and then again between 10,000 and 7200 ^{14}C yr BP.

We speculate that changes in fluvial geomorphic regimes influenced the geographic distribution of *Populus balsamifera* on the North Slope during the P–H transition. In the boreal forest, *P. balsamifera* is an early successional species that colonizes recently disturbed soils (Payette, 1993). In Alaska, this species grows mainly on river bars, where it forms large stands during the first 150 years of primary succession (Viereck, 1970; Viereck and Little, 1975). In the Arctic Foothills, relict stands of *Populus* now occupy recently deposited fluvial terraces, where they reproduce largely by root sprouting (Murray, 1980; Edwards and Dunwiddie, 1985). As described below, periods of stream aggradation in the Arctic Foothills occurred between 12,000 and 11,000 ^{14}C yr BP and again between 10,000 and 8700 ^{14}C yr BP. These two periods of aggradation created large areas of newly stabilized gravel bars suitable for *Populus* establishment. Newly deposited gravel bars were more rare during periods of channel incision. Perhaps the abundance and northerly extent of *P. balsamifera* varied according to the availability of its gravel-bar habitat.

5.3. Paludification of the Arctic Foothills and Arctic Coastal Plain

Dates on basal organic sediments from permafrost cores and stream cuts describe the timing of paludification in the study area (Appendix C). In general, two different types of basal organics occur in the tundra zone. The first are true basal organics (TBO), the lowest deposits of nonaquatic plant remains found in stratigraphic sections composed mainly of peat, which are immediately underlain by mineral sediments. The second are lowest cryoturbated organics (LCO) in permafrost, fragments of organic material that are now below the modern active layer. LCO is emplaced by cryoturbation, which churns organic matter deep into the active layer, where eventually it may become sequestered in permafrost (Ping et al., 1998). Ages of LCO samples are minimum-limiting dates on organic-matter accumulation at a site. We interpret TBO ages as absolute dates on the onset of organic accumulation.

To describe the paludification history of a topographically and pedologically representative portion of the Arctic Foothills, 19 TBO and LCO dates were obtained from a 10-km^2 area around the Mesa archaeological site (Kunz and Reanier, 1994). Auger sites were positioned along toposequences extending from rocky, frost-disturbed interfluves to alluvial toeslopes. The oldest dates range from 11,000 to 12,000 ^{14}C yr BP and come from TBO samples beneath midslope water tracks and toeslopes (Fig. 5, Appendix C). Soils on the interfluve southeast of the Mesa began storing carbon as early as 8400 ^{14}C yr BP, the AMS-radiocarbon age of a birch twig within a clump of LCO. Today this site is covered by moist acidic tundra growing among ice-wedge polygons. The wide scatter in the ages of ^{14}C-dated LCOs from convex slopes and interfluves probably results from the churning of silty soils there by frost heaving. Gravel prevented drilling deeper than 1–2 m at most of these upland sites. Stratigraphic sections provide additional information on the timing of paludification in the Arctic Foothills (Appendix C). The oldest TBO samples date to $\sim$12,600 ^{14}C yr BP and occur on toeslopes in headwater stream valleys.

Data collected by other workers elsewhere on the North Slope yield further insights into the regional time course of paludification. In compiling these data, we

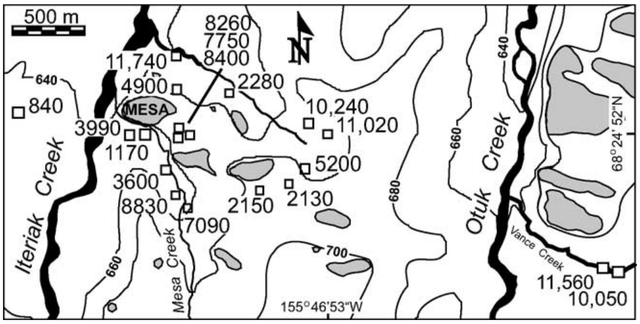

Fig. 5. AMS dates on basal organics from representative topographic positions near the Mesa archaeological site. The gray patches are the locations of either bedrock outcrops or the gravel surfaces of early Pleistocene stream terraces.

excluded dates on basal organics from sites where geomorphic evidence suggested a priori that the age of the underlying landform was $<14{,}000$ ^{14}C yr BP, as for example, Holocene-aged beach ridges and stream terraces. The goal is to identify when climate and soil conditions allowed paludification to occur. Hence we made exceptions to the 14,000-yr rule in the cases of sand dunes, where paludification could have caused dune stabilization, and at sites affected by solifluction lobes that overlie peat horizons (see below). Excluded from Appendix C are any dates on basal organics that accumulated in thaw lake basins (see Murton, 1996 for the identification of such sediments), dates whose 1-sigma errors are >200 yr, and dates from uncertain stratigraphic contexts.

Dates on TBOs from elsewhere on the North Slope (Appendix C) suggest that widespread peat accumulation occurred between $\sim 12{,}500$ and 11,000 ^{14}C yr BP. Frequencies of both TBO and LCO dates decline between 11,000 and 10,000 ^{14}C yr BP, probably reflecting a slowdown in paludification. Numerous TBO and LCO samples date to the interval 10,000–9000 ^{14}C yr BP, after which their numbers decline again. Together, basal-peat dates from the North Slope imply that primary productivity first surpassed decomposition $\sim 12{,}500$ ^{14}C yr BP, but only in topographic low points. Paludification slowed during the YD Chronozone, then resumed after 10,000 ^{14}C yr BP. By ~ 8500 ^{14}C yr BP, organic surface horizons probably had assumed their present wide distribution across the region.

5.4. Solifluction in the Arctic Foothills at the P–H boundary

At least one widespread episode of increased solifluction, the slow downslope movement of water-saturated sediments resulting from the thaw of frozen ground, affected the Arctic Foothills during the P–H transition. Today in arctic regions, solifluction accounts for a significant portion of slope denudation (Rapp and Åkerman, 1993), and it is an important supplier of sediment to streams. In stratigraphic sections, solifluction deposits appear as silty diamictons with elongate clasts that mostly dip downslope at 10–20° (Nelson, 1985; Elliott and Worsley, 1999). In the Arctic Foothills, solifluction sections are exposed most commonly in stream cutbanks within colluvial basins. Typically in these sections, a layer of deformed peat near the base of the section is overlain by solifluction deposits above an erosional contact. Undisturbed Holocene-aged peats, which often are several meters thick, overlie the solifluction deposits.

We obtained bracketing ^{14}C dates on solifluction deposits in four sections (Fig. 6, Appendix B). At Cobra Gulch, wood mixed with silty diamicton dates to 9740 ± 70 ^{14}C yr BP, and an overlying, horizontally bedded peat dates to 9670 ± 70 ^{14}C yr BP. In three other sections, solifluction silts are bracketed between dates of 12,340–8450, 10,430–8820, and 11,560–10,050 ^{14}C yr BP. At two other sites, the deformed peats underlying solifluction deposits dated to 11,130 (Fig. 6) and 11,350 ^{14}C yr (East Fork Etivluk River), but no upper limiting dates were found. It is possible that more than one solifluction episode occurred.

In the Arctic, increased solifluction occurs when summer thaw penetrates deeper than usual and melts ice lenses that accumulated previously at the base of the long-term, average active layer (Åkerman, 1993). Solifluction can also be enhanced by increased snow cover, which creates higher soil moisture upon melting (Matthews et al., 1993; Elliott and Worsley, 1999). Deeper thawing and increased winter precipitation at the YD/Holocene transition, and possibly during the earlier warm periods of the Lateglacial, are the likely causes of one or more widespread episodes of solifluction in the Arctic Foothills.

5.5. Fluvial history

Striking changes in erosional regime and channel planforms occurred in the streams draining the Arctic Foothills between 13,000 and 8000 ^{14}C yr BP. From at least 12,200 to $\sim 11{,}000$ ^{14}C yr BP, streams had braided channels and were aggrading rapidly. Talus buried part of the Lateglacial floodplain of Mesa Creek, a small meandering stream today that heads north of the Brooks Range outside the LGM glacial limit (Fig. 7). Channel fills are preserved at the talus/stream-gravel contact and contain organic debris dating to 11,400 ^{14}C yr BP (Appendix B), indicating that a braided stream had filled the valley with gravel by that time. Similarly, a fill terrace along the East Fork of the Etivluk River is capped by gravel interbedded with peat layers dating to 11,600–11,300 ^{14}C yr BP (Fig. 8). At section N1, the Nigu River had downcut through a valley fill of glacial outwash to near its present level before 12,200 ^{14}C yr BP

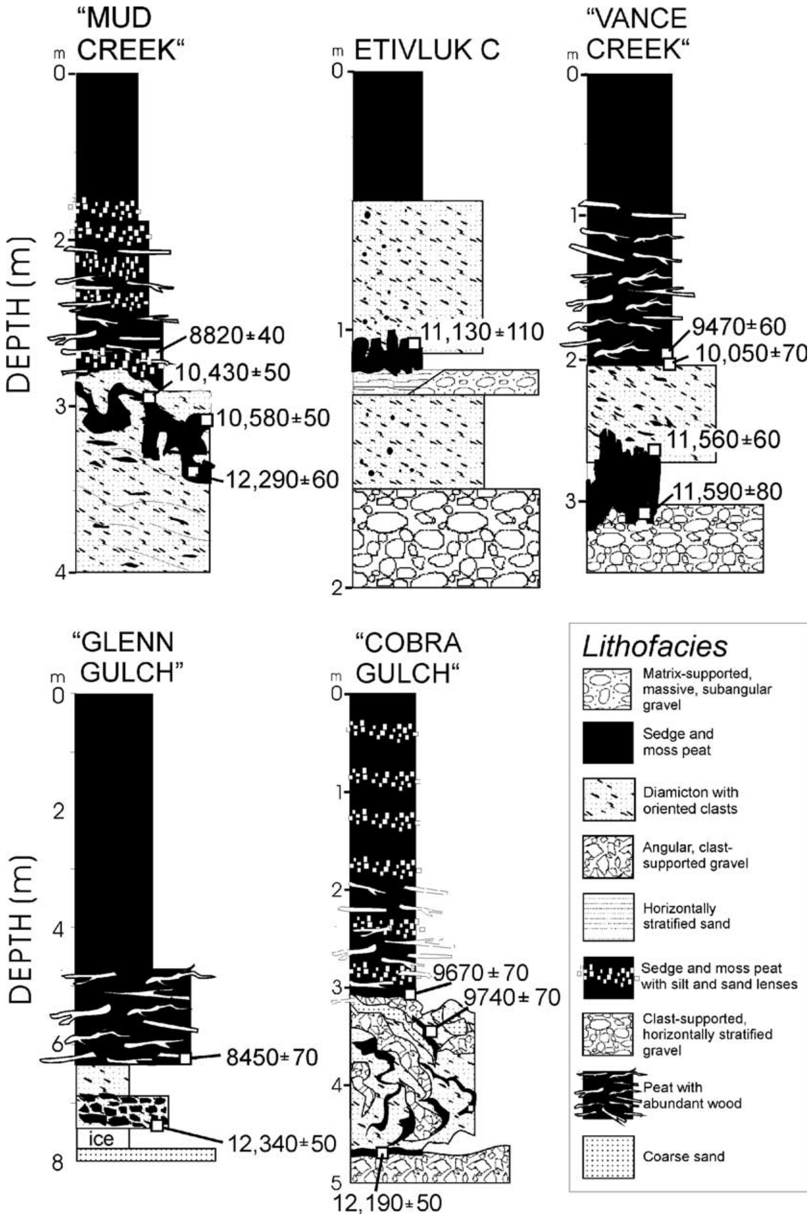

Fig. 6. Stratigraphic evidence for one or more episodes of solifluction that overrode Lateglacial peats around the time of the P–H transition.

(Fig. 9). Rapid deposition of gravel containing abundant organic lenses then occurred until $\sim$11,000 ^{14}C yr BP. At section N2 on the Nigu River, this same period of alluviation is represented by 10 m of sand interbedded with abundant organic debris. The age of a rooted willow buried in overbank sediments indicates that aggradation ceased there $\sim$10,900 ^{14}C yr BP.

The upper Ikpikpuk River valley contains well-exposed stratigraphic records of this stream's responses to Lateglacial and early Holocene climatic changes. Organic preservation is exceptional in this low-gradient, sandy stream, and much of the abundant organic debris encountered in sections there is reworked from older, frozen sediments (Nelson and Carter, 1987). For this reason, radiocarbon dates on water-transported plant debris provide maximum-limiting ages only. The most secure age control over fluvial history comes from dates on rooted plants buried in growth position and on aquatic macrophytes found in situ within oxbow-lake sediments.

Rapid alluviation was underway in the Ikpikpuk valley before 11,900 ^{14}C yr BP and continued until after 11,500 ^{14} yr BP, as indicated by the ages of willows buried in growth position several meters below the top of Terrace I at Little Supreme Bluff (LSB) (Fig. 10). Comparisons between channel orientations (accretion surfaces) and bedform orientations at LSB suggest that the Ikpikpuk River was braided during this aggradation interval. Bedforms record flow that was oriented $<60^\circ$ from the strike of associated channel axes, indicating the prevalence of downstream accretion, which is typical of

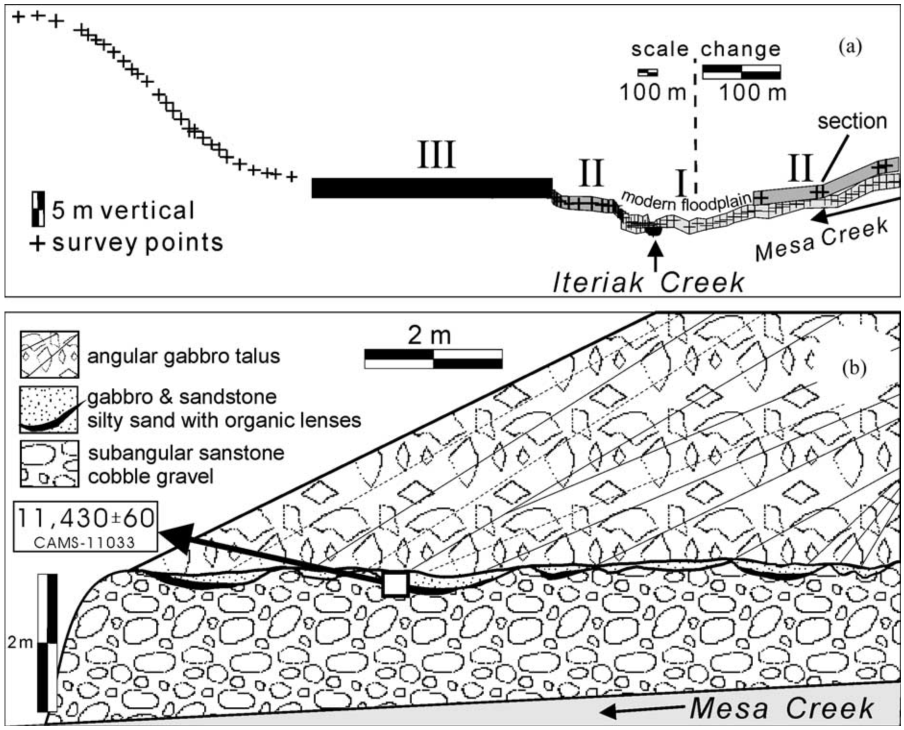

Fig. 7. Leveling transect parallel to lower Mesa Creek and across Iteriak Creek. A Lateglacial fill terrace (II) occurs along both creeks. The higher terrace (III) along Iteriak Creek probably dates to the Last Glacial Maximum. Surface I is the modern floodplain; (b). The surface of a Late Glacial floodplain is preserved at the contact between coarse, sandstone, fluvial gravels and overlying gabbro talus on the southern slope of the Mesa. The AMS-^{14}C date is from sedge seeds and moss fragments within a channel-fill deposit.

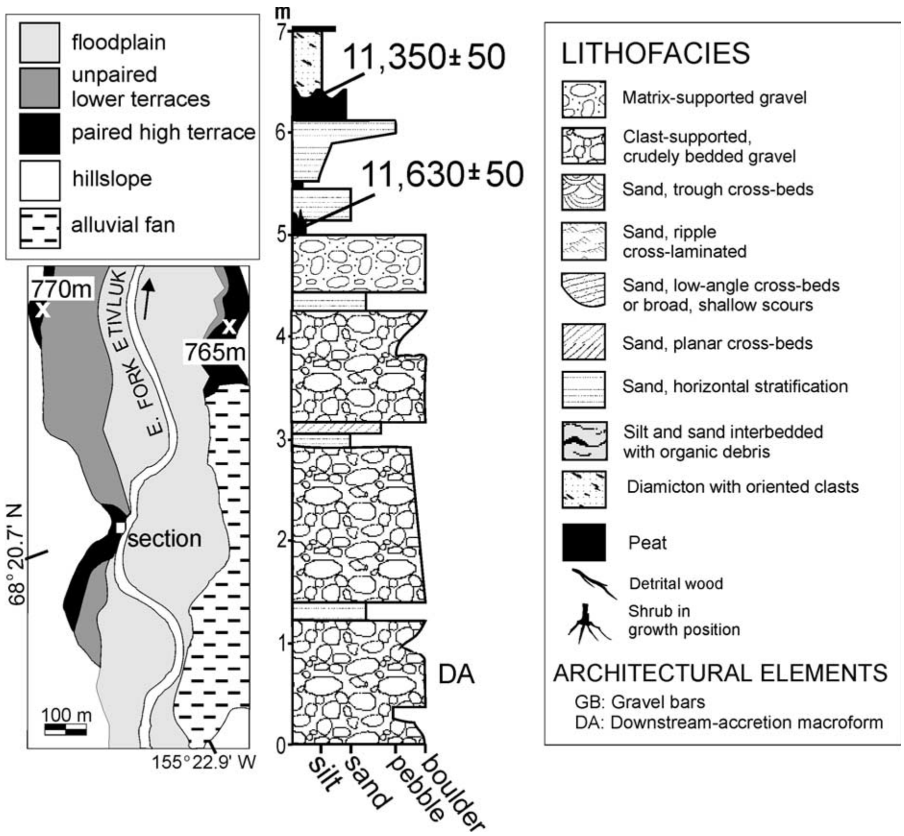

Fig. 8. Section through a fill terrace along the East Fork of the Etivluk River at the northern front of the Brooks Range. The lithofacies key follows Miall (1996).

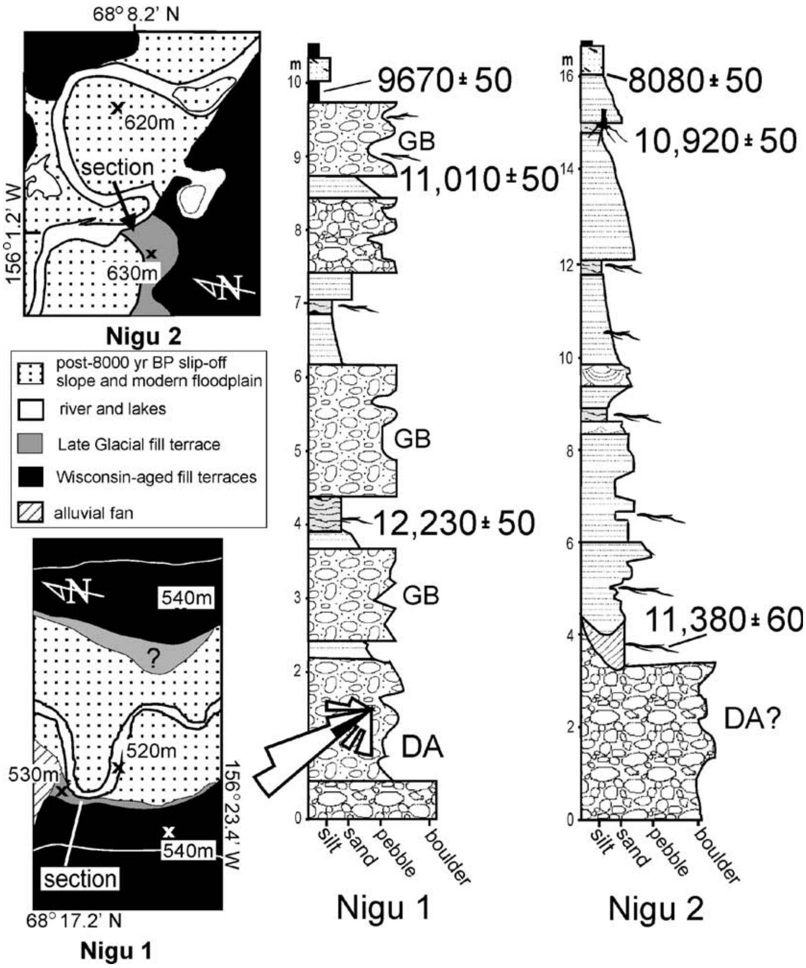

Fig. 9. Sections exposed along the Nigu River. Glacial-age fill terraces are composed of coarse, inorganic sediments and are associated with ice-marginal channels, moraines, and kames. The Lateglacial fill terrace is finer grained and contains abundant organic material. The uppermost ^{14}C date in each section is from capping peat. The rose diagram in the Nigu 1 diagram compares the dip directions of accretion surfaces (black) with flow directions inferred from imbricated cobbles (white). DA: downstream accretion macroforms probably deposited during ice retreat from the Last Glacial Maximum. GB: gravel bar of Lateglacial age.

braided channels (Miall, 1996). Only limited areas of Terraces I and II are preserved in the Ikpikpuk valley, making it difficult to infer channel planforms from channel scars on terrace surfaces (e.g., Huisink, 1997).

The Ikpikpuk River began a period of incision sometime before 10,900 ^{14}C yr BP, which took the river back to within several meters of its present altitude and produced an erosional unconformity marked by a lag of turf blocks. Similar turf blocks, which are quarried from tundra vegetation along cutbanks, are common in the channel today (Fig. 11). The ancient turf blocks at LSB date to 11,700–11,600 ^{14}C yr BP, but these dates imply only that channel incision and accompanying bluff erosion occurred sometime afterwards. A closer limiting date comes from downstream at Cottonwood Bend. Here a similar channel-lag, also exposed deep inside Terrace II, produced a date of 10,900 ^{14}C yr BP and indicates that incision occurred after that time (Fig. 12).

At Cottonwood Bend, aggradation resumed after 9700–9100 ^{14}C yr BP, maximum-limiting dates on plant macrofossils contained in the upper 4 m of this bluff (Fig. 13) (Nelson and Carter, 1987). The Ikpikpuk valley was refilled with sandy sediments to pre-YD levels before 8800–8700 ^{14}C yr BP, the ages of emergent, aquatic macrophytes found in the basal sediments of oxbow lakes on Terrace II at both LSB and Cottonwood Bend (Figs. 10 and 12). At LSB, comparisons between channel (i.e., accretion surfaces) and bedform orientations suggest that the Ikpikpuk River was meandering during this early Holocene aggradation interval. In many parts of the section, bedforms record flow oriented $>60°$ from the strike of associated channel axes, indicating the predominance of lateral accretion typical of meandering channels (Miall, 1996) (Fig. 10).

Descending series of unpaired terraces in the Arctic Foothills indicate net downcutting after $\sim$8700 ^{14}C yr BP. Along the Ikpikpuk River, willow shrubs buried in growth positions by fluvial sands indicate that the floodplain was 6–7 m above its present level between 6070 (site IK3) and 5430 ^{14}C yr BP (site IK1; Fig. 1).

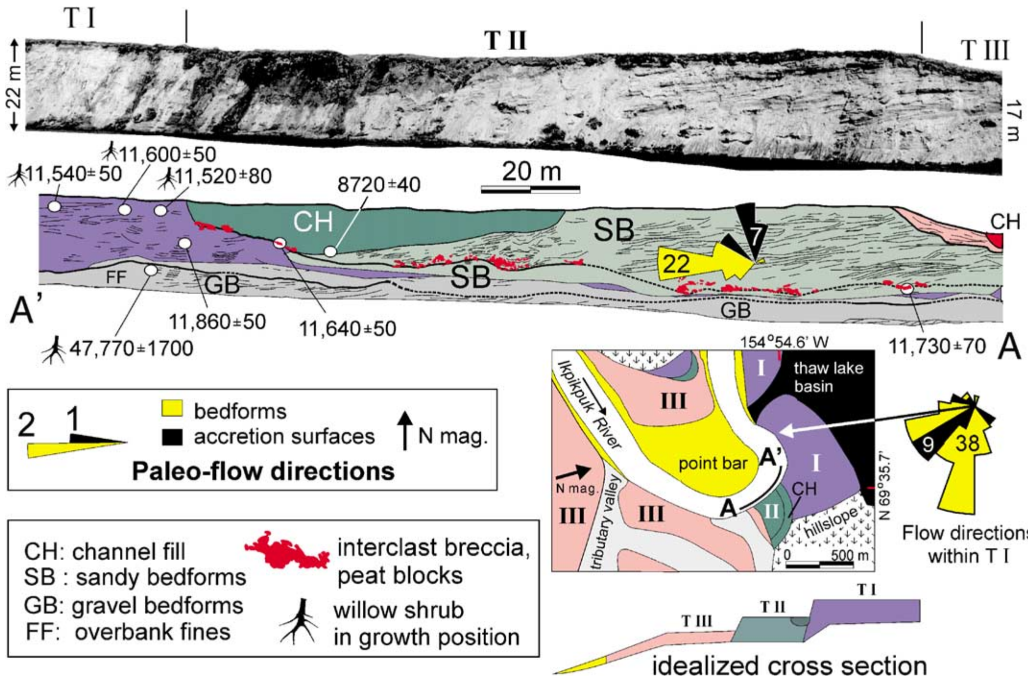

Fig. 10. Photo-mosaic and stratigraphic interpretation of the central part of Little Supreme Bluff, Ikpikpuk River. Different colors in the diagram represent sedimentary units separated by fifth-order bounding surfaces (Miall, 1996). Rose diagrams show dip of accretion surfaces (black) and paleocurrent directions (yellow). DA: downstream-accretion macroforms, LA: lateral-accretion macroforms, FF: floodplain fines, SB: sandy bedforms, GB: gravel bedforms, CH: channel. Terrace I is remnant from a valley fill emplaced before $\sim$11,000 ^{14}C yr BP, probably by braided channels. Terrace II is a remnant from a second valley fill deposited by meandering channels in the early Holocene. The 8720-yr date comes from seeds of the emergent macrophyte, *Menyanthes trifoliata*, contained in laminated, silty organics at the base of the infilled channel on Terrace II. *Menyanthes* grows today in oxbow lakes isolated from river flooding. Rose diagrams compare the orientations of small-scale bedforms with larger accretion surfaces.

Fig. 11. A lag of peat blocks in ancient channel deposits at Cottonwood Bend, Ikpikpuk River. White lines trace third-order bounding surfaces between lateral-accretion macroforms. Paleo-flow was to the right in direction of peat-block imbrication. A 10,940-yr ^{14}C date came from within the circled area. The entrenching tool is 60 cm long.

Similarly, at 5010 ^{14}C yr BP, Iteriak Creek (site IT1) was depositing gravel over willows rooted 2 m above its modern floodplain.

In summary, fluvial stratigraphy suggests that streams in the Arctic Foothills incised to near their present altitudes sometime before 12,200 ^{14}C yr BP (Fig. 14). Rapid alluviation occurred from at least 12,200 to $\sim$11,000 ^{14}C yr BP and was accompanied by braided channels in the Mesa Creek valley and in the Ikpikpuk valley at Little Supreme Bluff. Between $\sim$11,000 and $\sim$10,000 ^{14}C yr BP, downcutting occurred. After 10,000 ^{14}C yr BP, valley aggradation resumed and was accompanied by meandering channel planforms near Little Supreme Bluff. Net downcutting began after 8700 ^{14}C yr BP that has continued to the present. These changes occurred in catchments regardless of whether their headwaters were in the Brooks Range or in the Arctic Foothills, eliminating the possibility that glaciers controlled their dynamics (cf., Ashley and Hamilton, 1993; Hamilton, 2001). The fact that the Ikpikpuk River was downcutting during the YD, a time when the Ikpikpuk Dunes were active on the Arctic Coastal Plain (Carter, 1993), indicates that changes in base level caused by dune movements downstream (cf., Loope et al., 1995) did not affect fluvial dynamics at the study sections.

6. Synthesis

Data describing multiple biological and geological processes are needed to reconstruct the history of a landscape and its component ecosystems. Besides yielding a better-rounded picture of the paleoenvironment, this approach enables the insensitivities, nonlinear

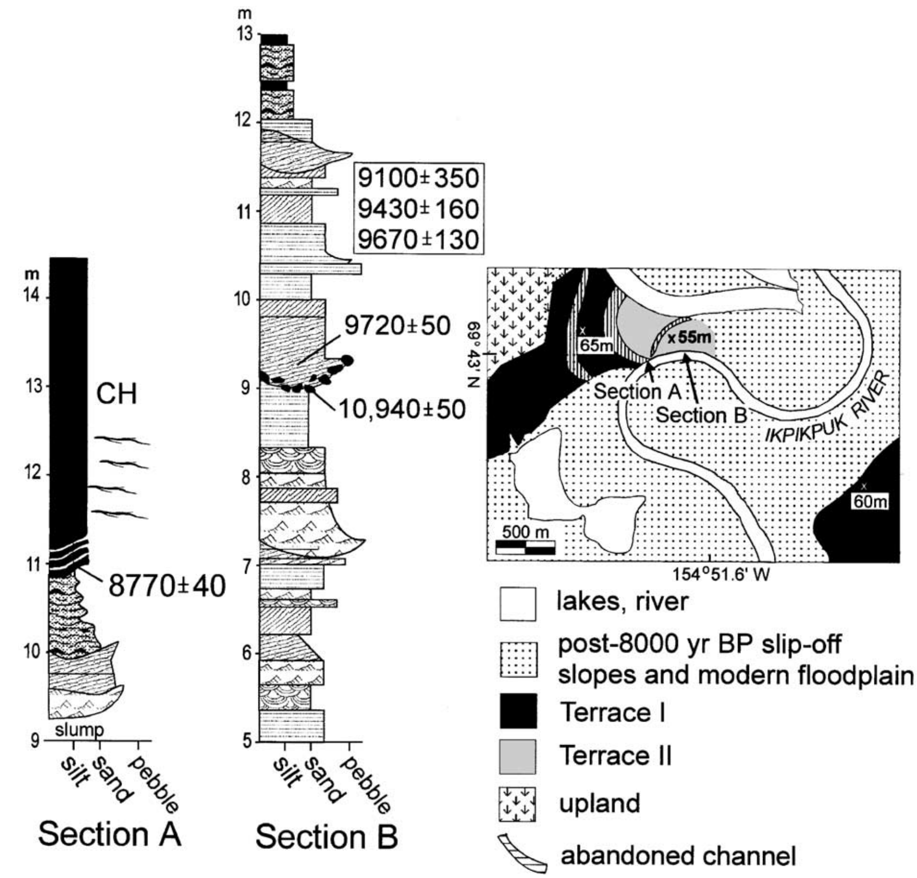

Fig. 12. Stratigraphics sections at Cottonwood Bend, Ikpikpuk River. The two prominent fill terraces here are correlative to Terraces I and II at Little Supreme Bluff based on relative altitude, cross-cutting relationships, and limiting ^{14}C ages. The ^{14}C date in Section A is from *Menyanthes trifoliata* seeds in silty, graminoid peat at base of an oxbow lake on Terrace II. The boxed ^{14}C dates in Section B are the ages of detrital organics reported by Nelson and Carter (1987). In Section B, the 10,950-yr date is the age of a single poplar leaf from sand infilling the channel-lag shown in Fig. 11. The 9720-yr date is the age of another single poplar leaf from the overlying channel sands.

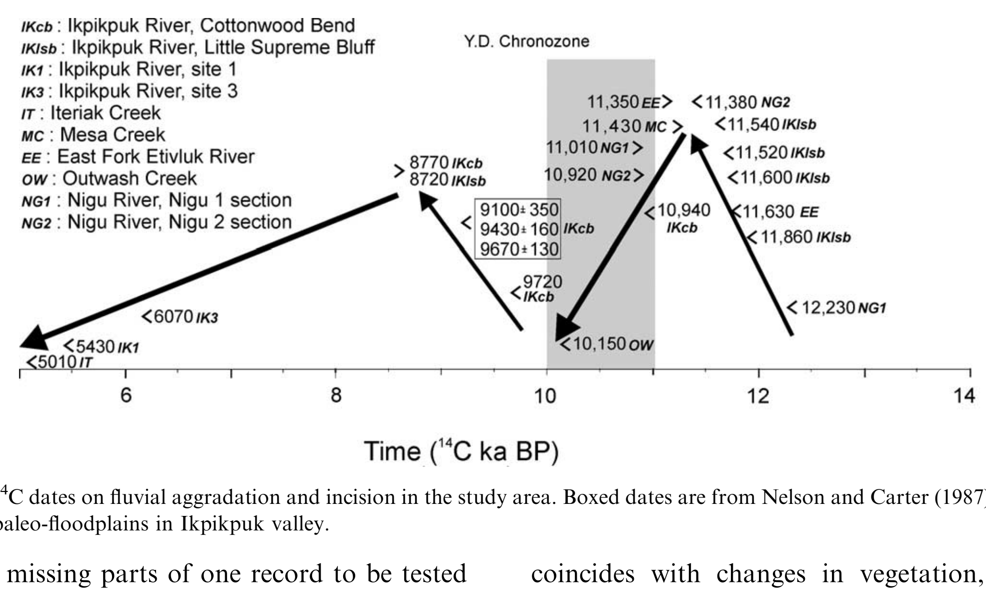

Fig. 13. Limiting ^{14}C dates on fluvial aggradation and incision in the study area. Boxed dates are from Nelson and Carter (1987). The vertical axis is relative height of paleo-floodplains in Ikpikpuk valley.

responses, and missing parts of one record to be tested and corrected by the intact parts of others.

The common theme in the paleo-records described here from the P–H transition in the Arctic Foothills is the importance of changes in moisture. The history of lake-level fluctuations indicated by the LOP sections coincides with changes in vegetation, organic-matter accumulation, floodplain dynamics, and solifluction in the surrounding region (Fig. 14). Rising water levels in LOP around 12,500 ^{14}C yr BP marked the beginning of a period of rapid alluviation by braided streams which lasted until $\sim$11,000 ^{14}C yr BP (Fig. 14). Increased

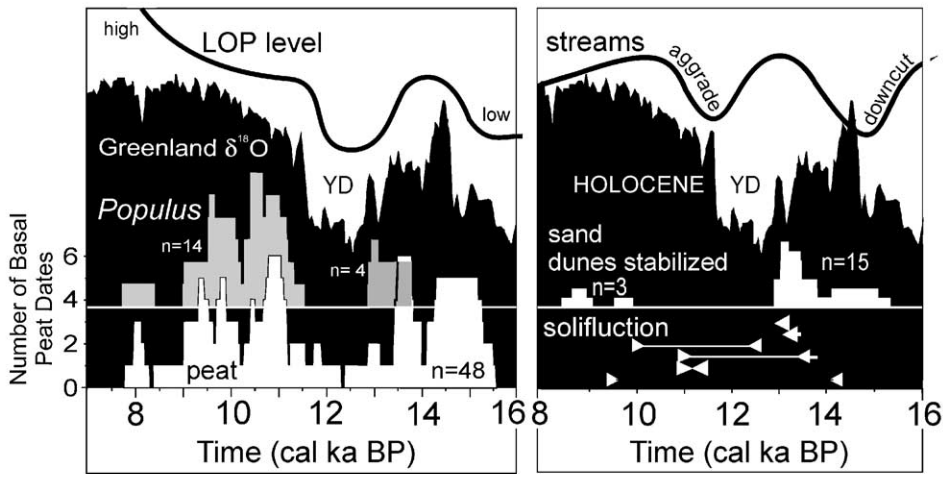

Fig. 14. Synthesis of environmental changes in the Arctic Foothills during the P–H transition. The time scale is in calendar years to allow comparison with the Greenland δ^{18}O record (Grootes and Stuiver, 1997). The generalized lake-level history is inferred from stratigraphy at LOP. The peat dates are from Appendix C, and the *Populus* dates from Appendix A. A similar history for *Populus* emerges from LOP pollen data. The histograms depict the number of ^{14}C dates whose calibrated 1L age range falls within a given decade. Dates on stabilization of the Ikpikpuk Dunes are from Carter (1993).

alluviation probably was caused by a combination of increased permafrost melting and heightened hillslope erosion, which perhaps was triggered by increased summer rainfall on a landscape with discontinuous vegetation cover (cf., Cogley and McCann, 1976; Edlund et al., 1989).

Rising water levels in LOP ~12,5000 ^{14}C yr BP also coincide with the spread of shrub tundra in the Arctic Foothills as indicated by pollen records from LOP and Tukuto Lake (Oswald et al., 1999). At the same time, peat deposition began all across the North Slope (Fig. 14), probably also in response to increasing effective moisture, a well-known trigger for paludification (Moore, 1987; Ovenden, 1990; Gorham, 1991). Prior to 10,000 ^{14}C yr BP, peat deposition was limited to topographic low points in the Arctic Foothills, suggesting that effective moisture was lower than during the Holocene. *Populus* trees expanded their range between 12,000 and 11,000 ^{14}C yr BP in response to warmer summers and the wider availability of recently deposited alluvium. The pre-YD occurrence of *Populus* trees in the Artic Foothills is indicated by ^{14}C dates on wood and a leaf from the Nigu and Ikpikpuk valleys and by the occurence of *Populus* pollen in the LOP1 section.

Elsewhere in northern and interior Alaska, there also is evidence for a major increase in effective moisture ~12,500 ^{14}C yr BP. Glaciers in the Brooks Range underwent a minor re-advance between ~13,000 and 11,500 ^{14}C yr BP, possibly in response to increasing winter snowfall (Hamilton, 1986). On the Arctic Coastal Plain, the Ikpikpuk dunes became inactive between ~12,500 and 11,000 ^{14}C yr BP, most likely in response to increased soil moisture (Carter, 1993). During this same interval in interior Alaska, wetter conditions led to the erosion of extensive gully systems on loess slopes (Hamilton et al., 1988), and water levels at Birch Lake rose more than 18 m between 12,700 and 12,200 ^{14}C yr BP (Abbott et al., 2000).

During the YD, water levels fell in LOP and paludification slowed in the Arctic Foothills (Fig. 14). Streams incised their floodplains, probably because declining precipitation reduced the input of slope sediments (Fig. 14). Sediment input also may have declined because of reduced solifluction activity and fewer landslides caused by permafrost melting. *Populus* retracted its distribution in the Arctic Foothills, perhaps in response to cooler summers and to the shrinkage of its floodplain habitat as streams became entrenched. On the Arctic Coastal Plain, dunes were reactivated after 11,000 ^{14}C yr BP in response to decreased soil moisture (Carter, 1993).

Around 10,000 ^{14}C yr BP water levels in LOP again rose, and *Populus* again expanded its distribution in the Arctic Foothills. A brief episode of widespread solifluction occurred, probably in response to the melting of ground ice that had accumulated during the YD. Rapid alluviation resumed in valleys, and this time, at least in the Ikpikpuk valley, channels assumed meandering planforms. This second aggradation episode probably was triggered by intense slope erosion caused by increased summer rains, deeper thawing of soils, and widespread solifluction. In northwestern Canada, numerous thaw lakes developed between 10,000 and 9000 ^{14}C yr BP) in response to an increase in active-layer thickness (Burn et al., 1986; Burn, 1997. Temporary northward extensions of thermophilic plant taxa on the North Slope and in northwestern Canada suggest that summer temperatures were higher than today during the earliest Holocene (Ritchie et al., 1983; Nelson and Carter, 1987; Anderson, 1988).

The LOP sections suggest that water levels rose further after $\sim$8600 ^{14}C yr BP. This is consistent with findings elsewhere in the region. On the Arctic Coastal Plain, increasing soil moisture again stabilized the Ikpikpuk dunes $\sim$8500 ^{14}C yr BP (Carter, 1993). In interior Alaska, water levels at Birch Lake rose markedly between 8800 and 8000 ^{14}C yr BP (Abbott et al., 2000; Edwards et al., 2001).

The LOP pollen record suggests that the modern vegetation of the Arctic Foothills, along with its poorly drained, peaty soils, probably was established between 9000 and 8500 ^{14}C yr BP. Other pollen records from the North Slope (Eisner and Peterson, 1998; Oswald et al., 1999) are consistent with this interpretation. The widespread establishment of organic soil horizons was a major turning point for ecosystem history in the Arctic Foothills. Soil temperatures must have fallen and soil moisture risen because of the insulating and water-holding properties of the organic horizons (Bockheim et al., 1998). These same organic surface horizons restricted the frost heaving of mineral material and reduced soil erosion, thus depriving streams of sediment inputs and forcing them into a long-term trend of floodplain incision. As floodplains narrowed and became vegetated, the rate of loess deposition in downwind areas slowed, probably enhancing soil acidification and promoting further paludification (Walker and Everett, 1991).

It is obvious that many natural processes in the Arctic are controlled by temperature. It is less obvious that moisture plays a crucial role in determining the impacts of climatic change there. Moisture is important in mediating the impacts of climatic change in the Arctic for several reasons, some of which are unique to high latitudes. First, atmospheric circulation patterns and the low saturation vapor pressure of cold air restrict precipitation in polar regions (Walsh et al., 1994). Second, the amount and phase state of water at the often-frozen tundra surface determine surface energy budgets to a greater extent than they do at lower latitudes (Kane, 1996). Third, permafrost and seasonally frozen soils create a hydrologic setting where the availability of liquid water is seasonally limited (Hinzman et al., 1996). Consequently, small changes in the timing of thaw or of liquid precipitation can have large effects on land surface processes and vegetation. Fourth, the presence of permafrost makes paludification possible within this relatively dry environment, and paludification triggers powerful feedbacks between soil moisture, soil temperature, soil disturbance, and vegetation cover (Zoltai and Tarnocai, 1975; Moore, 1987). Finally, fluvial activity controls the areal extent of the active floodplains where loess originates. Loess influences soil genesis, vegetation, and possibly herbivore populations in downwind areas (Walker and Everett, 1991; Walker et al., 2001).

The inherent sensitivity of arctic ecosystems to changing moisture conditions, combined with the large shifts in water balance that occurred during the P–H transition, made water a key factor in determining the impacts of climate change on landscapes in the Arctic Foothills. During the last glacial maximum, northern Alaska was much drier than today because atmospheric circulation patterns were more zonally oriented, the Polar Front was shifted further south, and lowered sea levels greatly increased climatic continentality (Hamilton et al., 1993; Abbott et al., 2000; Edwards et al., 2001; Guthrie, 2001). Effective precipitation increased rapidly as postglacial sea-levels flooded the broad continental shelves of central Beringia and brought moist, cloudy, maritime climates back to the region. Our results indicate that changes in moisture were the proximate causes for ecosystem-wide responses to climate change in the Arctic Foothills at the end of the last ice age and that they will be again in the future.

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Appendix A

Radiocarbon ages of extralimital *Populus* subfossils in northern Alaska and northwestern Canada are given in Table 1.

Appendix B

Radiocarbon ages from lake of the pleistocene, solifluction sections, and fluvial sections described in this study (Table 2).

Appendix C

Radiocarbon ages of basal peats and buried organics on the North Slope (Table 3).

Table 1

Laboratory number	Location (Lat (°N), Long (°W))	Dated material	δ ^{13}C (‰)	^{13}C-adjusted radiocarbon age (yr before 1950 AD)	1σ calibrated age range (cal yr before 1950 AD) [a]
Ikpikpuk River (this study)					
Beta-109676[b]	"Cottonwood Bend" 69° 43′, 154° 52′	Leaf of *Populus balsamifera*	−29.2	10,940±50[b]	13,100–12,880
Beta-111032[b]	"Cottonwood Bend" 69° 47′, 154° 48′	Leaf of *Populus balsamifera*	−31.8	9720±50[b]	11,200–11,120
Beta-120018	"Dune Bend" 69° 43′, 154° 52′	*Populus* log	−26.9	9350±60	10,670–10,430
Beta-121113	"Cottonwood Bend" 69° 43′, 154° 53′	*Populus* log	−29.6	9270±60	10,560–10,290
Ikpikpuk River (Nelson and Carter, 1987)					
I-11280	69° 48.9′, 154° 24.73′	Peat containing *Populus balsamifera* leaves	—	9540±160	11,170–10,580
I-13174	69° 43′, 154° 53′	*Populus* wood	—	9430±160	11,070–10,428
I-11282	69° 35.8′, 154° 54.5′	*Populus* wood	—	9380±150	11,040–10,290
I-13324	69° 17.5′, 154° 42.5′	*Populus* wood	—	8710±140	10,110–9540
Northern Alaska and Northwestern Canada (Hopkins et al., 1981)					
GSC-1514	Twin lakes near Inuvik, Northwest Territories	*Populus* wood	—	11,500±160	13,800–13,180
W-1254	East shore of Kotzebue Sound, southeast of Riley's Wreck	*Populus* wood	—	11,340±400	13,810–12,970
I-10274	West Bank of Nigu River, 11 km north of Inyorurak Pass (probably this paper's Nigu 1 section)	*Populus* wood	—	11,100±170	13,180–12,910
GSC-2022	Coastal bluffs southeast of Sabine Point, Yukon Territory	*Populus* wood	—	9940±90	11,550–11,230
1-11073	lkpikpuk River, Arctic Foothills	*Populus* wood	—	9670±130	11,200–10,750
W-2620	Seward Peninsula between Rex Point and Cape Deceit	*Populus* wood	—	9625±350	11,340–10,430
W-1255	East side of Cape Blossom, Kotzebue Sound	*Populus* wood	—	9020±400	10,670–9550
W-1249	East shore of Kotzebue Sound at Arctic Circle	*Populus* wood	—	8550±400	10,160–9030
W-1993	Sagavanirtok River valley, northern Alaska	*Populus* wood	—	8400±300	9680–9010
W-1250	North side of Cape Blossom, Kotzebue Sound	*Populus* wood	—	7270±350	8410–7740

[a] Using Calib4 (Stuiver et al., 1998).
[b] AMS date.

Table 2

Sample/Location[a]	Laboratory number	Location Lat (N), Long (W)	Dated material	$\delta^{13}C$ (‰)	^{13}C-adjusted radiocarbon age (yr before 1950 AD)	1σ callibrated age range (cal yr before 1950 AD)[b]
Ikpikpuk River						
7-12-97a, Howard Hill	Beta-111035	69° 25′, 154° 47′	Willow in growth position[c]	−30.7	5430±60	6290–6120
7-16-99a, middle reaches	Beta-132694	69° 39′, 154° 50′	Willow in growth position	−26.1	6070±80	7140–6760
7-14-99Q1, "Cottonwood Bend"	Beta-132693	69° 43′, 154° 532′	Seeds of *Menyanthes trifoliata*	−27.1	8770±40	9910–9690
Ikpikpop1, "Cottonwood Bend"	Beta-109676	69° 43′, 154° 52′	Poplar leaf	−29.2	10,940±50	13,100–12,880
Ikpikpop2, "Cottonwood Bend"	Beta-111032		Poplar leaf	−31.8	9720±50	11,200–11,120
7-13-99a, Little Supreme Bluff	Beta-132631	69° 36′, 154° 55′	Seeds of *Menyanthes trifoliata*	−25.4	8720±40	9880–9560
7-14-99d, Little Supreme Bluff	Beta-132692	69° 36′, 154° 55′	Willow in growth position[c]	−27.3	11,520±80	13,790–13,410
7-14-99a, Little Supreme Bluff	Beta-132632	69° 36′, 154° 55′	Willow in growth position[c]	−28.3	11,540±50	13,790–13,430
7-1-98c, Little Supreme Bluff	Beta-120023	69° 36′, 154° 55′	Willow in growth position[c]	−26.4	11,600±50	13,800–13,450
7-1-98a, Little Supreme Bluff	Beta-120022	69° 36′, 154° 55′	Stick cf willow[c]	−27.4	11,640±50	13,820–13,460
7-1-98q, Little Supreme Bluff	Beta-120024	69° 36′, 154° 55′	Stick cf willow	−29.1	11,730±70	13,840–13,500
7-1-98x, Little Supreme Bluff	Beta-120025	69° 36′, 154° 55′	Stick cf willow	−27.8	11,860±50	14,040–13,660
7-18-98c, Little Supreme Bluff	Beta-123316	69° 36′, 154° 55′	Willow in growth position[c]	−29.0	47,770±1700	—
"*Mesa Creek*"						
July93d1	Beta-69896	68° 25′, 155° 48′	Sedge seeds, moss	−34.6	11,430±60	13,760–13,180
Nigu River						
24June98f, Section 1	Beta-120020	68° 8′, 156° 1′	Sedge stem	−26.8	8080±50	9030–9000
24June98e, Section 1	Beta-121110		Willow in growth position[b]	−26.1	10,920±50	13,010–12,880
24June98a, Section 1	Beta-120019		Stick cf. willow[b]	−30.1	11,380±60	13,460–13,170
9June97kin-d, Section 2	Beta-124782	68° 17′, 156° 24′	Twig cf birch	−27.5	9670±50	11,170–10,890
9June97kinC, Section 2	Beta-111127		Stick cf. willow[b]	−28.0	11,010±50	13,140–12,900
7-13-98b, Section 2	Beta-120537		Stick cf. willow[b]	−29.2	12,230±50	15,230–14,110
Iteriak Creek						
7-22-98h	Beta-121111	68° 36′, 155° 46′	Stick cf. willow[b]	−28.2	5010±80	5890–5650
East Fork Etivluk River						
17July97b1	Beta-111036	68° 21′, 155° 23′	Sedge seeds	−28.2	11,350±50	13,440–13,160
17July97b5	Beta-108257		Twig cf. birch	−27.7	11,630±50	13,810–13,460
"*Lake of the Pleistocene*"						
LOP1-j	Beta-97512	68° 36′, 156° 16′	Twig of birch or willow	−30.2	4900±60	5660–5590
LOP1-f	Beta-97511		Twig of birch or willow	−30.2	8360±70	9470–9280
LOP1-e	Beta-97510		Twig of birch or willow	−30.5	8530±70	9450–9490
LOP1-d	Beta-97509		Twig of birch or willow	−29.0	8870±60	10,150–9790
7-21-98d3	Beta-122124		Twig of birch or willow	−28.0	9020±50	10,220–10,180
LOP1-c	Beta-97508		Twig of birch or willow	−30.5	9570±60	11,110–10,700
7-21-98d3	Beta-121105		Twig of birch or willow	−28.8	10,240±50	12,300–11,760
7-21-98c2	Beta-120539		Twig of birch or willow	−29.2	10,270±50	12,320–11,770

Table 2 (*continued*)

Sample/Location[a]	Laboratory number	Location		Dated material	$\delta^{13}C$ (‰)	^{13}C-adjusted radiocarbon age (yr before 1950 AD)	1σ callibrated age range (cal yr before 1950 AD)[b]
		Lat (N)	Long (W)				
7-21-98a2	Beta-123317			Twig of birch or willow	−27.3	10,280±50	12,330–11,780
7-21-98ax	Beta-120536			Twig of birch or willow	−29.8	10,300± 40	12,340–11,930
LOPabB	Beta-111033			Twig of birch or willow	−29.6	10,350±40	12,360–11,960
LOP1d4	Beta-125086			Twig of birch or willow	−29.1	10,380±40	12,600–11,970
7-21-98b2	Beta-122123			Twig of birch or willow	−28.0	10,400±50	12,623–11,980
6July97Brown	Beta-108004			Twig of birch or willow	−29.7	10,470±50	12,800–12,190
LOP1-b	Beta-97507			Twig of birch or willow	−31.9	10,720±70	12,930–12,640
7-21-98ax3	Beta-121106			Twig of birch or willow	−29.0	10,950±50	13,110–12,890
LOP1-a	Beta-97506			Twig of birch or willow	−30.9	12,320±60	15,300–14,140
7-21-98ax1	Beta-122125			Twig of birch or willow	−28.1	12,450±60	15,390–14,260
6July97z	Beta-108006			Twig of birch or willow	−27.6	13,420±50	16,350–15,890
6July97,85–90	Beta-108264			Twig of birch or willow	−27.7	34,920±260	—
6July97,80–85	Beta-108263			Twig of birch or willow	−27.9	29,570±150	—
9July97x	Beta-108261			Twig of birch or willow	−27.8	40,650±530	—
6July97,55–60	Beta-108262			Twig of birch or willow	−30.0	43,780±780	—
6July97,50–55	Beta-108005			Twig of birch or willow	−21.0	47,830±1300	—
Solifluction sections							
"Vance Creek" D	Beta-67362	68° 25′,	155° 44′	Twig of willow or birch	−28.1	9470±60	11,040–10,580
"Vance Creek" AH	Beta-69888			Twig of willow or birch	−29.0	10,050±70	11,900–11,260
"Vance Creek" B	Beta-69903			Twig of willow or birch	−25.8	11,560±60	13,800–13,430
"Vance Creek" A	Beta-69902			Twig of willow or birch	−26.2	11,590±80	13,810–13,440
"Mud Creek" G	Beta-133727	68° 26′,	156° 26′	Willow stick[c]	−25.3	8820±40	
"Mud Creek" F	Beta-103629			Willow stick	−25.0	10,430±50	12,769–12,140
"Mud Creek" E	Beta-103628			Willow stick	−29.9	10,580±50	12,850–12,370
"Mud Creek" D	Beta-103627			Willow stick	−28.4	12,290±60	15,280–14,130
3July93c1, Etivluk River	Beta-69887	68° 35′,	156° 16′	Sedge seeds and stems	−28.1	11,130±110	13,180–13,000
"Cobra Gulch" B	Beta-107861	68° 26′,	155° 57′	Willow stick[c]	−28.3	9670±70	11,180–10,810
"Cobra Gulch" D	Beta-107862			Willow stick[c]	−26.9	9740±70	11,200–11,140
"Cobra Gulch" A	Beta-111126			Twig cf. birch	−28.0	12,190±50	15,190–14,100
"Glenn Gulch," 7-22-98c	Beta-120540	68° 27′,	155° 38′	Willow stick	−26.1	8450±70	9530–9330
"Glenn Gulch," 7-22-98b	Beta-121108			Twig of willow or birch	−26.8	12,340±50	15,310–14,140

[a] Informal place names in quotes.
[b] Using Calib4 (Stuiver et al., 1998).
[c] Dated by conventional radiometric techniques using liquid scintillation. Others by the AMS technique.

Table 3

Sample / Lowest basal peat (TBO) or lowest cryoturbated peat in permafrost (LCO)[a]	Laboratory number	Location		Dated material	$\delta^{13}C$ (‰)	^{13}C-adjusted radiocarbon age (yr before 1950 AD)[b]	1σ callibrated age range (cal yr before 1950 AD)[c]
		Lat (N)	Long (W)				
Arctic Foothills (This study)							
Wolf Creek[d] A–TBO	Beta-103635	68° 31′ 10″,	155°36′	Sedge stem, Wood fragment	−28.3	12,640±70	15,510–14,380
Otuk Tributary[d] —TBO	Beta-121108	68° 27′,	155°38′ 10″	Twig	−26.8	12,340±50	15,310–14,140
Mud Creek[d] —TBO	Beta-103627	68° 26′,	156° 28′	Willow leaf	−28.4	12,290±60	15,280–14,130
Cobra Gulch[d] —TBO	Beta-111126	68° 26′ 9″,	155° 56′ 50″	Willow branch	−27.6	12,190±50	15,190–14,100
2July96j—LCO	Beta-101361	68° 25′ 5″,	155° 47′ 52″	Wood fragment	−27.3	11,740±60	13,840–13,510
Vance Creek[d] —TBO	Beta-69902	68°24′ 30″	155°36′ 30″	Twig	−26.2	11,590±80	13,810–13,440
Nigu C—TBO	Beta-69887	68° 34′ 48″,	156° 16′	Sedge stem	−28.1	11,130±110	13,180–13,000

Table 3 (*continued*)

Sample / Lowest basal peat (TBO) or lowest cryoturbated peat in permafrost (LCO)[a]	Laboratory number	Location	Dated material	$\delta^{13}C$ (‰)	^{13}C-adjusted radiocarbon age (yr before 1950 AD)[b]	1σ callibrated age range (cal yr before 1950 AD)[c]
		Lat (N) Long (W)				
2July96e—LCO	Beta-101357	68° 24′ 52″, 155° 46′ 53″	Wood fragment	−27.2	11,020±60	13,140–12,900
2July96f—LCO	Beta-101358	68° 24′ 53″, 155° 46′ 59″	Sedge stem	−28.5	10,240±60	12,310–11,760
4July93a (Vance Creek[d] auger hole)—TBO	Beta-69888	68° 24′ 30″, 155° 36′ 30′	Twig	−29.0	10,050±70	11,900–11,260
Cobra Gulch[d] —TBO	Beta-107861	68° 26′ 9″, 155° 56′ 50″	Twig	−28.3	9670±70	11,180–10,810
Vance Creek[d] —TBO	Beta-67362	68° 24′ 30″, 155° 36′ 30″	Twig	−28.1	9470±60	11,040–10,580
Grayling Gulch[d] —TBO	Beta-103632	68° 34′ 48″, 156° 16′	Birch twig	−30.4	9310±50	10,580–10,420
14July93b—LCO	Beta-69891	68° 24′, 155° 48′ 3″	Twig	−26.7	8830±100	10,150–9700
Otuk Tributary[d] —TBO	Beta-120540	68° 27′, 155° 38′ 10″	Twig	−26.1	8450±70	9530–9330
16July93d—LCO	Beta-69894	68° 24′ 33″, 155° 47′ 10″	Twig	−28.3	8400±60	9490–9330
16July93a—LCO	Beta-69892	68° 24′ 33″, 155° 47′ 10″	Sedge stem	−27.9	8260±90	9430–9030
16July93b – LCO	Beta-69893	68° 24′ 33″, 155° 47′ 10″	Sedge stem	−27.6	7750±140	8640–8390
Wolf Creek[d] B—TBO	Beta-103626	68° 31′ 10″, 155° 36′	Sedge stem	−25.0	7210± 50	8110–7960
14July93c—LCO	Beta-67360	68° 24′ 40″, 155° 48′ 7″	Twig	−27.2	7090±80	7970–7790
2July96c—LCO	Beta-101355	68° 24′ 36″, 155° 46′ 27″	*Betula* twig	−28.9	5200±60	5990–5910
2July96h—LCO	Beta-101359	68° 25′, 155° 47′ 49″	Twig	−25.9	4900±50	5660–5590
2July96L—LCO	Beta-101363	68° 24″ 41″, 155° 47′ 60″	*Betula* bark	−29.2	3990±50	4520–4410
14July93aa—LCO	Beta-69890	68° 24′, 155° 48′ 3″	Sedge stem	−30.2	3600±60	3980–3780
2July96d—LCO	Beta-101356	68° 24′ 36″, 155° 46′ 27″	*Betula* twig	−28.3	3320±50	3630–3470
2July96i—LCO	Beta-101360	68° 24′ 58″, 155° 47′ 23″	*Betula* twig	−29.2	2280±60	2350–2160
2July96m—LCO	Beta-101364	68° 29′ 36″, 155° 46′ 51″	*Betula* twig	−30.8	2150±50	2300–2060
2July96b—LCO	Beta-101354	68° 24′36″, 155° 46′36″	*Betula* twig	−27.9	2130±50	2290–2010
2July96k—LCO	Beta-101362	68° 24′ 40″, 155° 48′ 7″	*Betula* twig	−29.8	1170±60	1170–990
North Slope (Previous studies)						
Hamilton (1979)—TBO	USGS-163	68°8′, 147° 11′	Wood	—	9600± 85	11,160–10,700
Wind River						
Hamilton (1982)	I-10715	68° 49′, 149° 17.5′	Sedge peat	—	9460±150	11,090–10,430
Toolik River—TBO			Peat	—	11,700±180	14,000–13,450
East Fork Etivlik River—TBO	I-11,419	—	Peat	—	11,500±140	13,800–13,190
Hinzman et al. (1991)	—	—				
Imnavait Creek—TBO						
Carter (1993)	I-11600	69° 1′ 18″, 151° 56′ 42″	Wood	—	11,800±170	14,060–13,500
Ikpikpuk Dunes—TBO						
—TBO	I-12177	70° 25′ 42″, 152° 34′ 14″	Peat	—	11,700±180	14,000–13,450
—TBO	USGS-448	70° 22.4′, 153° 12′ 12″	Peat	—	8180±75	9270–9020
—TBO	I-10887	70° 22′ 24″, 153° 12′ 12″	Peat	—	8660±150	9890–9530
—TBO	I-11420	70° 22′ 24″, 153° 12′ 12″	Peat	—	8010±130	9060–8640
—LCO	Beta-5384	70° 4′ 21″, 151° 22′ 54″	Peat	—	9330±90	10,670–10,290
—LCO	Beta-5383	70° 4′ 21″, 151° 22′ 54″	Peat	—	9600±100	11,170–10,700
Everett and Brown (1982)	DIC-706	—	Organic matter	—	7330±95	8280–7980
Sagwon—LCO			Over gravel			
—LCO	DIC-707	—	organic matter over gravel	—	8860±125	10,180–9710
Ping, C.-L. (written communication, 1999)	GX-20889	70° 16″	Peat	−29.2	7300±70	8180–7980
Arctic Coastal Plain—LCO						
Schell and Zieman (1983)	—	70° 31″, 149° 6′	Peat	−28.3	8430 est. 200	9630–9090
—TBO						
—TBO	—	70° 33′, 149° 24′	Peat	−28.7	9050 est. 200	10,500–9790
—TBO	I-6839	70° 31′, 149° 19′	Peat	—	12,610 est. 200	15,560–14,290
—TBO	I-6838	70° 31′, 149° 52′	Peat	—	8550 est. 200	9890–9300
‘Arctic Coastal Plain’—TBO	—	—	Peat	−27.2	9805 est. 200	11,560–10,760

[a] TBO and LCO terms explained in text.
[b] All ages in this table are AMS dates except Beta-111126.
[c] Using Calib4 (Stuiver et al., 1998).
[d] Informal place name.

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